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Acta Miner. Petr., Szeged

AN ORE-GENETICAL STUDY OF PYRRHOTITE FROM MT. CSÁKÁNYKŐ

A. Embey Isztin

INTRODUCTION

Several years ago, in the pyroxenic andesite quarry of Mt. Csákánykő, Mátra Mountains, North Hungary, inclusion-like nests of pyrrhotite, attaining even 4—5 cm in diametre, were found. The study of this peculiar, rather rare occurrence (with a view to the geological and petrographical conditions of the region) was a tentative for settling the problem of the genesis of these interesting minerals and of their eventual connection with the other ore mineralizations of the region.

GEOLOGICAL AND PETROGRAPHICAL CONDITIONS OF THE REGION

Belonging to the eastern part of the Mátra Mountains, Mt. Csákánykő lies at a distance of about 3 air kilometres south of Recsk. The inclusion-dotted rock was encountered during quarry operations.

In this region the most ancient rocks (red and grey limestone, clay and siliceous shale) can be referred to the Ladinian stage of the Triassic.

The diabases of the Miklós valley, which petrographically coincide with the spilitized vesicular diabases of Mt. Darnó at Sirok, are of Middle Triassic age (or possibly Cretaceous). The Oligocene is only represented by Rupelian clay-marls.

The Miocene formations are more significant in the eastern marginal zone of the Mátra Mountains. On the one hand, they are represented by sedimentary desposits (sandstone, sandy clay, schlier); on the other hand, by various volcanic rocks of considerable thickness (rhyolitic tuff, dacitic tuff, andesitic tuff, and pyroxenic andesite).

The phytofossiliferous sandstones, referred to the Aquitanian or Burdigalian stages, are overlain by variegated clays and conglomerates. Finally, the Burdigalian ends with the so-called Lower Rhyolite Tuffs.

The volcanic complex, making up the Mátra Mountains, was formed in Late Helvetian and Tortonian times. Initially, eruptions of andesitic, dacitic, and rhyolitic tuffs took place and after that the andesite mantle dikes were formed. As suggested by I. Kubovics [1963], the chronological order of the eruptions of the Helveto—Tortonian volcanic complex was the following:

TABLE 1

Chronological Order of Helveto-Tortonian Eruptions

Serial number	Formation	Stage
1.	Andesite mantle and dikes	
2.	Middle Andesite Tuff II Biotitic Dacitic Tuff Middle Andesite Tuff I	Tortonian
3.	Middle Rhyolite Tuff	
4.	Lower Andesite Tuff	Helvetian

The latest volcanic products and the andesite dikes of funnel or broadening laccolith shape show hardly any difference in mineralogical composition. The main

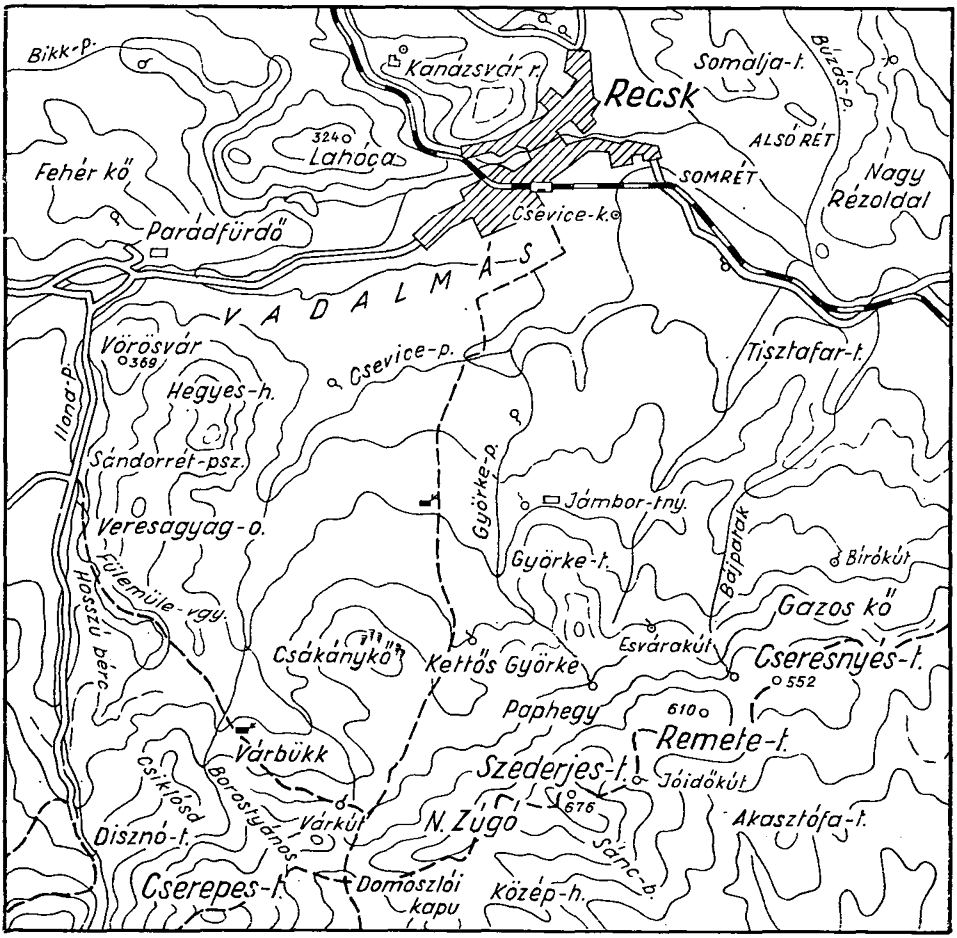


Fig. 1. Map scheme of environs of Csákánykő

difference between the two kinds of rock is in their texture: the degree of crystallization of the matrix. Geologically and petrographically, the andesite dikes can be included in two groups: the Györke dikes and the Farkashegy dikes. The first group is characterized by a development at subvolcanic depth. As for mineralogical composition, the Györke dikes are of carbonate character, whereas this cannot be observed either in the Farkashegy dikes or in the rock of the andesite mantle.

The Csákánykő andesite belongs to the Györke dikes, its carbonate content is considerable (3—5% on the average). The more so, large cavities or pockets (nests), containing crystallized carbonate minerals, are also frequent. As shown by the measurements of B. NAGY [1967], the carbonate content of the andesite is very high close to the pockets. This indicates that the carbonate-bearing pockets must have been formed by gradual saturation. According to B. NAGY [1967], the sequence of carbonate mineralization was the following: spherosiderite → siderite → ankerite → dolomite → calcite → aragonite. This succession indicates an incrase in Ca with decreasing temperature. The source for carbonatization is believed to have been provided by the melting of great amounts of carbonate rock. The Csákánykő andesite, itself, was formed in Helvetian schlier, and GY. VARGA (personal communication) found various types of foreign xenoliths in it.

The lack of carbonates in certain dikes and in the andesite mantle as well as their varying abundance in other dikes seems to be due to differences in pressure during the consolidation of the magma. A comparatively higher pressure — such as would be required for the formation of carbonates — can be supposed to have existed at the subvolcanic level only. Because of its high partial pressure, CO_2 will escape from the magma, if the pressure due to the overburden is low or zero.

Thus the carbonatization of the Csákánykő andesite suggests a consolidation under the conditions of considerable overburden pressure. Accordingly, the degree of crystallization of the matrix is high. The average size of the matrix feldspars varies between 100 and 20 μ. The mineralogical components of the andesite have shown the following percentage distribution:

Porphyric impregnation:		
	plagioclase	33,2%
	hypersthene	10,1%
	carbonate	4,3%
	opaque (mainly ilmenite)	1,6%
Matrix:		
	plagioclase	40,5%
	augite	10,3%
	Total:	100,0%

Most abundant among the porphyric constituents, plagioclase is usually of labradoritic-bytownitic composition; the nucleus of the largest porphyric plagioclases is mainly bytownitic, and even a plagioclase of An = 90% composition was measured. The most common twin laws are the albite, albite-Carlsbad, and albite-Ala laws. The frequency order is only approximate.

Over the majority of plagioclases a multiple zoning can be observed.

The second, significant porphyric component is hypersthene. For the most part, it shows parallel extinction, though some zonation in inner of the crystals can also be observed. This suggests the incorporation of some Ca. The augitization of the margin of hypersthene indicates changes which must have occurred during crystallization.

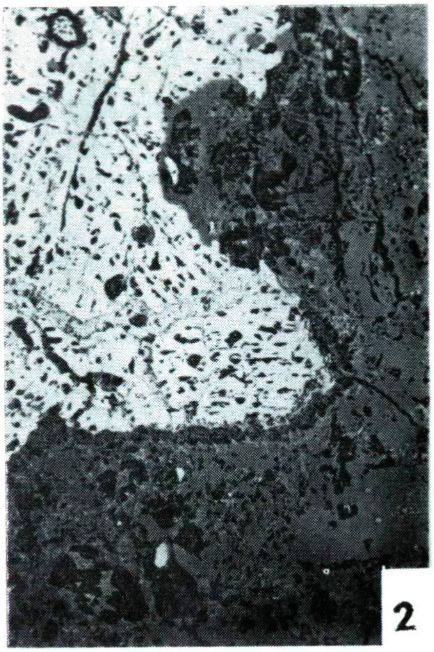
The mafic constituent of the matrix is chiefly augite-pyroxene. The carbonate inclusions appear in the form of radial nests. The nests are of completely irregular shape. In the neighbourhood of carbonates, some chalcedony also occurs. Close to the porphyric constituents, carbonate is abundant. The shape and environment of the carbonate nests allow to conclude that the carbonates have resulted from the displacement of the matrix before the rock was completely consolidated. The Ca of the displaced matrix feldspars was bound as $CaCO_3$. Thus the alkali content of the feldspar laths relatively increased, i. e. their composition was shifted toward albite. With increasing alteration, the feldspar was completely decomposed, and chalcedony precipitated. A comparatively higher volatile content may considerably decrease the final temperature of consolidation of the rock. For this reason, the carbonatization of the matrix took place as late as the hydrothermal phase. As already mentioned, the Csákánykő hypersthene andesite contains plenty of xenoliths. Of these, we have to deal with the crystalline quartz inclusions, too. The rare pyrrhotite inclusions are mostly associated with these quartz inclusions. However, both the volume and the abundance of the quartz inclusions are much greater than those of pyrrhotite. At the same time, some pyrrhotite can be observed in the andesite even independently of the quartz inclusions.

The quartz inclusions are of hypidiomorphic character, their grain size averages 0,6 to 0,3 mm. They have an elongated form, and the combination of $(10\bar{1}0)$ and $(10\bar{1}1)$ can sometimes be observed. The fine cracks of the quartz inclusions and the interstices have been filled later by carbonates. Occasionally, it can be observed that both the quartz inclusions and the adjacent andesite are traversed by a carbonate streak. This seems to suggest that while entering the magma the silica was not accompanied by carbonate, but when later the andesite was carbonatized, the solution impregnated the quartz inclusions, too. As observed by the author, on the margin of the quartz crystals (characterized, otherwise, by totally faultless optics), at their contact with the andesite, an undulated, imperfect extinction is obtained and the resolution and recrystallization of the quartz are associated phenomena. This anomalous extinction is due to crystallization-induced volume increase and to the resultant tension stresses.

The contact between the quartz inclusions and the andesite is locally sharp, in other places a wreath-shaped diopside-augite rim of 0,50 to 0,35 mm width can be observed.

The quartz and the various xenoliths derive from the basement, their source rocks being not known sufficiently as yet. All that which seems to be probable is that, on the basis of its textural and morphological characteristics, the quartz must have been formed as a low-temperature α-quartz in a late-crystallization phase of the magma. This suggestion is supported by B. NAGY's X-ray analyses, too (personal communication).

Fig. 2. Pyrrhotite coating the porphyric feldspars. ‖N, ×33
Fig. 3. Pyrrhotite coating the porphyric feldspars. ‖N, ×33
Fig. 4. Quartz inclusions in pyrrhotite, ‖N, ×160
Fig. 5. Oriented quartz inclusions. ‖N, ×160



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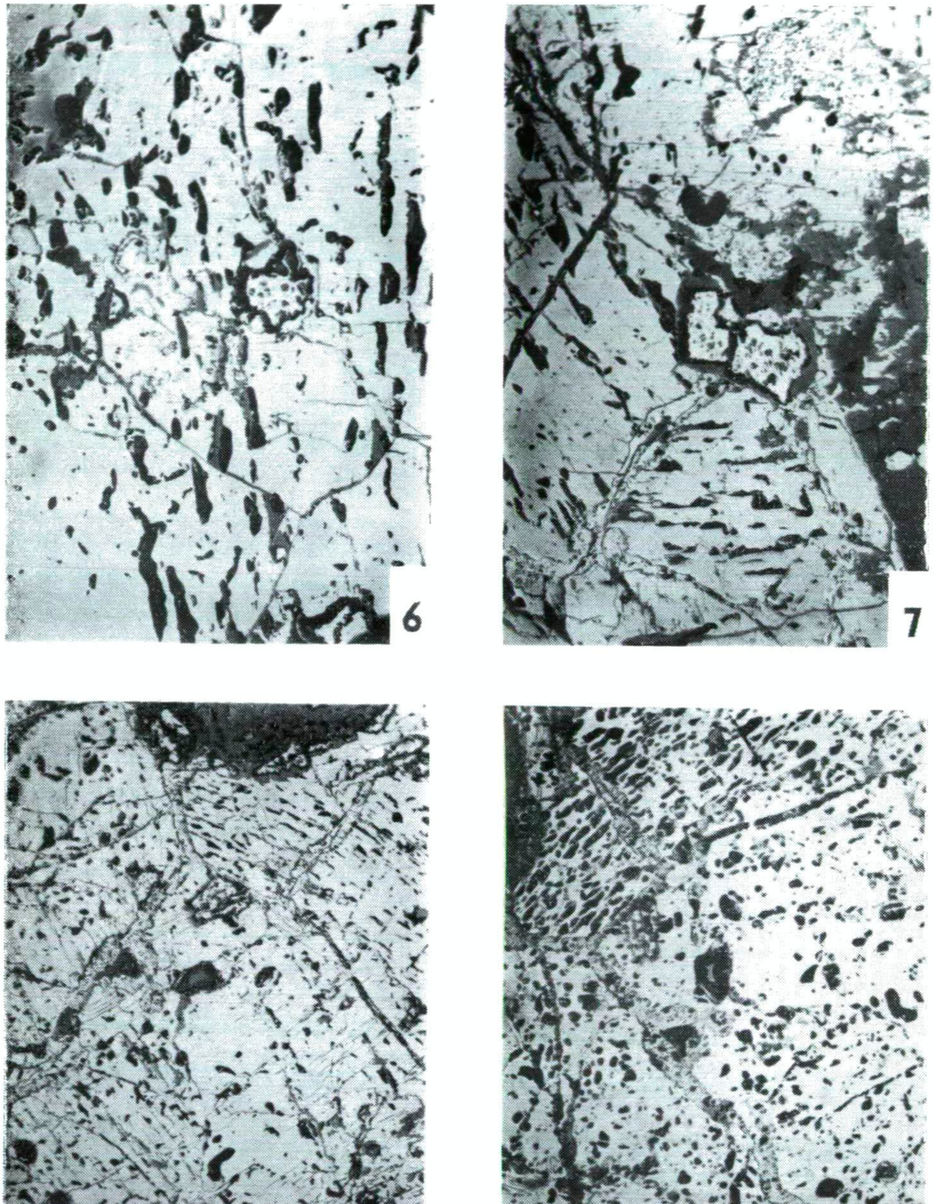
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Pyrrhotite occurs in the form of minor nests, accumulations or of thin crusts, often close to — or even within — the quartz inclusions. In such cases, it appears along grain contacts or fills tiny cracks. However, it may sometimes appear in the andesite independently of the quartz inclusions. In such cases, the porphyric plagioclases of the andesite are coated by thin pyrrhotite films which have quasi protected the morphological and geometric characteristics of the crystal (Plate I, *Fig. 2, 3*).

Pyrrhotite is poorly crystallized, massive inclusion material. As already mentioned, the association of quartz inclusions and pyrrhotite is not compulsory. Hence, it is probable that there is no genetic relationship between them.

Under ore microscope, pyrrhotite is allotriomorphic, granular, sections according to the (0001) being rare. The predominant grain size is 2,5 to 0,5 mm. Its mean reflectivity corresponds to that of the Kisbánya (Gutin Mountains, Rumania) and Nagybörzsöny pyrrhotites which were examined for the sake of comparison. (Measured with a Zeiss photoelectric instrument in a green light, all three gave a value of 40%.)

Pyrrhotite may locally contain idiomorphic, elongated, prismatic quartz inclusions, some of which show a controlled orientation (Plate I, *Fig. 4, 5*). Along fissures and joints, pyrrhotite is slightly pyritized. The pyrite grains are of irregular shape; sometimes incompletely developed hexahedral faces are observable (Plate II, *Fig. 6, 7*). Pyrite is one of the ore minerals which possess the highest energy of attachment. That is why it is represented by an idiomorphic crystallization, even though belonging to the later separation.

The amount of chalcopyrite found in pyrrhotite was considerably higher than that of pyrite. On the basis of the X-ray spectroscopic results, the Cu content of the ore is estimated at about 1%. Chalcopyrite chiefly occurs at the margins of the pyrrhotite grains or in the fine cracks traversing them (Plate II, *Fig. 8, 9*). It is relatively abundant in the vicinity of idiomorphic quartz inclusions, and even in their skeleton-crystal-like cavities. It may occasionally look like a "pseudomorph" after hexagonal quartz. The chalcopyrite formed later has partly replaced the pyrrhotite and even the pyrite. At higher magnification, the isolated remnants of the displaced pyrrhotite in the chalcopyrite can be well observed (Plate III, *Fig. 10*).

In a very small amount galena is also present in the ore inclusion. Its mostly poorly developed tiny crystals occur in the chalcopyrite veinlets traversing the pyrrhotite. The centre of the veinlets is filled by the last-precipitated carbonates. The more or less circular, carbonate-filled nests and pockets in the pyrrhotite exhibit the same arrangement. The pockets are ring-like covered by a thin chalcopyrite film, and on the inner side of the ring there are one or two hexahedral crystals of galena. Finally, the centre of the pocket is filled by carbonate (Plate III, *Fig. 11, 12, 13*). The galena, occurring on the margins of the pockets, is composed of greater but fewer grains than the galena in the veinlets. In the larger crystals of galena even cleavage could be observed.

Fig. 6. Pyrite grains of irregular shape. ||N, ×80
Fig. 7. Half-developed pyrite hexahedra. ||N, ×80
Fig. 8. Chalcopyrite streaks between pyrrhotite grains. ||N, ×33
Fig. 9. Chalcopyrite streaks between pyrrhotite grains. ||N, ×33

TABLE 2

X-ray powder data of the Csákánykő and Kisbánya pyrrhotites

Pyrrhotite Csákánykő		Pyrrhotite Kisbánya		Pyrrhotite ASTM/2-1341		Chalcopyrite ASTM/5-0490	
I	d_{hkl}	I	d_{hkl}	d_{hkl}	I	d_{hkl}	I
mst	3,03	w	3,04	—	—	3,03	100
st-mst	2,97	st-mst	2,97	2,98	60	—	—
st	2,635	st	2,643	2,64	80	—	—
vw	2,442	vw	2,450	2,45	10	—	—
w	2,265	w	2,263	2,26	10	—	—
vst	2,058	vst	2,055	2,06	100	—	—
w	1,887	w	1,884	1,88	10	—	—
vw	1,868	—	—	—	—	1,865	40
st	1,719	st	1,718	1,72	70	—	—
mst	1,603	mst	1,606	1,61	40	1,591	60
vvw	1,569	—	—	—	—	1,573	20
mst-d	1,424	mst	1,425	1,43	40	—	—
mst-st	1,320	mst-st	1,320	1,32	60	1,323	10

vst: very strong
st: strong
mst: middle strong
w: weak
vw: very weak
d: diffuse

FeKα, 40 kV, 8 mA, Exp: 5^h

The paragenetic sequence of the ore mineral association of Csákánykő was the following: quartz → pyrrhotite → pyrite → chalcopyrite → galena → carbonates.

The carbonate, which was formed last, seems to be identical with the carbonate which displaced the andesite matrix. The carbonatization of the andesite must have taken place under hydrothermal conditions, in the last stage of the magmatic process. This suggestion appears to be warranted by the paragenesis.

Practically, that is all which can be deduced from the observed characteristics of the ore minerals of "transient" character.

The examination of the minor element content of pyrrhotite suggests a hydrothermal origin.

The Co content of the Csákánykő pyrrhotite is about four times greater than the Ni content. In the liquid-magmatic pyrrhotite, connected with peridotite or norite, the Ni content exceeds that of Co. The highest value of Co is 1%, whereas that of Ni may even attain 8 to 10%. Consequently, in the liquid-magmatic minerals Co is always less abundant than Ni. Thus the liquid-magmatic pyrrhotite described by S. KOCH [1955] from Szarvaskő in the Bükk Mountains and characterized by a pentlanditic intergrowth, contained 28 700 ppm Ni and only strong traces of Co.

However, hydrothermal processes may enrich the residual Co of the magma in a greater measure. That is the reason why the hydrothermal iron minerals, e. g. pyrite, are dominated by Co, while their Ni is quite poor.

Fig. 10. Chalcopyrite displacing pyrrhotite. ||N, ×315

Fig. 11. Carbonate-filled cavity with a chalcopyrite ring at its margin and with crystals of galena on the inner side of the ring. ||N, ×33

Fig. 12. Hexahedral crystals of galena among chalcopyrite and carbonate grains. ||N, ×250

Fig. 13. Pyrrhotite, chalcopyrite, and galena at the margin of a carbonate-filled cavity. ||N, ×250.

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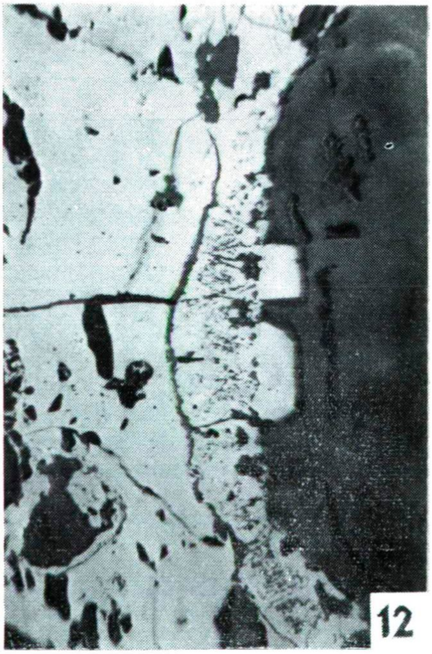

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In this case too, Co seems to be camouflaged by Fe. The bivalent Ni has such a small ionic radius that it can no longer enter the pyrrhotite under hydrothermal conditions. The abundance of Ni is also insufficient, thus being indetectable by X-ray analysis in an eventual separated phase.

Platinum metals cannot be detected by conventional spectroanalytical methods. However, B. NAGY (paper presented at the ordinary session of the Mineralogical Section of the Hungarian Geological Society), who used a method of enrichment, could detect a very low Pt content.

TABLE 3

Semiquantitative results (in ppm) of the spectral analyses of pyrrhotite, quartz grains and andesite from Csákánykő

	Pyrrhotite	Quartz grains	Andesite
Ag	—	1,6	—
B	25	10	<10
Ba	—	100	1000
Co	6000	20	60
Cr	—	6	60
Cu	1000	100	50
Ga	—	1	6
Li	<4000	—	1600
Mn	1000	2500	1600
Ni	<1600	16	40
Pb	—	160	25
Sr	250	1600	2500
Ti	6000	600	16000
V	—	<2,5	250
Zn	160	160	160

The origin of the Csákánykő pyrrhotite ore inclusion is characterized by assimilation process during which the assemblage of elements has changed. In the process the high volatile content of the magma must have played a decisive role, and the volatiles may have been largely enriched by the wet sedimentary environment on which the volcanic products were deposited. Because of the abundance of volatiles, the final consolidation (crystallization) of the magma took place, as already referred to by the author, at a considerably lower temperature (hypomagmatic character). Consequently, the rock was not yet consolidated when the volatile vapours percolated it, so that the volatiles themselves could be crystallized in cavities and pockets of varying size scattered throughout the rock body. Mineralized pockets can only be formed under the effect of a corresponding counter-pressure (damping effect). Consequently, they can develop at the subvolcanic level only. The source of the Csákánykő ore mineralizations seems to have had something in common with the near-by mineralizations. The ironpoor (enargitic) mineralization of the Mt. Lahóca lies rather close to the Csákánykő exposure. It can be supposed that the inclusions of the andesite derive from the basic-igneous basement of Bükk type. It is known that near Recsk a deep-seated pyrite-chalcopyrite mineralization was disclosed by deep drilling. As believed by the author, this deep subsurface process of great importance may be connected with the ore sources from which the pyrrhotite inclusions of the Csákánykő derive.

ACKNOWLEDGEMENT

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DATA ON THE GEOLOGY AND MINERALOGY OF THE EPLÉNY MANGANESE ORE DEPOSIT

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INTRODUCTION

True, the sedimentary manganese ore deposit of Epléný (Bakony Mts.) in its practical importance lags behind the Úrkút manganese ore deposit, however, in view of the complexity and great interest of genetical problems it deserves a more thorough, detailed study. As it appears from the number of references, there has been no lack of interest in this respect up to the present time.

The Epléný manganese ore became known in 1928 as a surface outcrop, somewhat later than the Úrkút one and at the very beginning there was merely surface mining, then gradually the subsurface mining was begun. A remarkable development in the manganese ore mining in Epléný set on after the second world war when between the years 1945 and 1966 deep-reaching borings and subsurface exposures

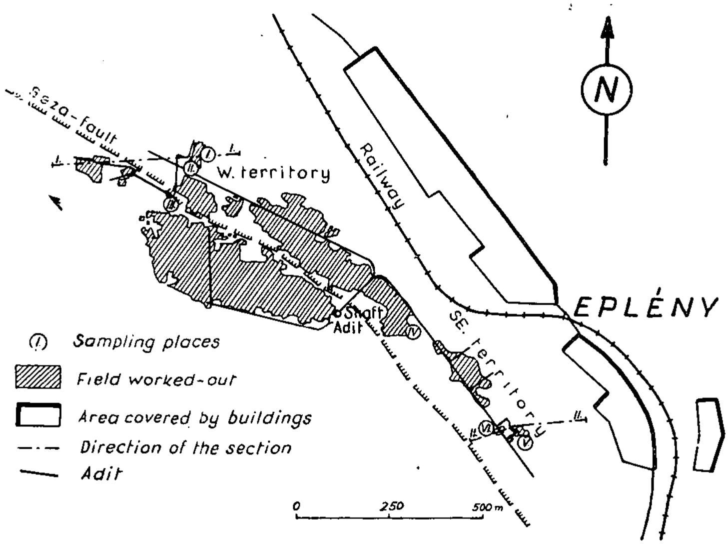

Fig. 1. Map scheme of the Epléný manganese ore deposit showing sampling places and direction of geologic sections

made possible the delimitation of the productive manganese ore beds and a considerable increase of production here.

Sampling places of the Epleny area, place and direction of sections shown in *Figs. 2, 5* and *6* are illustrated in the map to *Fig. 1.*

GEOLOGY OF THE AREA

The Eplény manganese ore occurrence is found in a trench-like structure surrounded by Triassic rocks. Within the area of the occurrence a varied Jurassic sequence of strata had developed. The development of Jurassic in Eplény, as compared with the area of the Úrkút manganese ore field, shows unconformity, what is obvious from the fact that proceeding from the Triassic borders towards the inner parts of the basin more and more complete layers can be found.

Jurassic

Lower Liassic. "Dachstein" limestone (white, light-grey, yellow, dense, finecrystalline) representing the uppermost series of Triassic turns with conformity into Liassic limestone of the Dachstein type with similar characteristics. The upper banks of the limestone formation are of oölitic texture, with Foraminifera, very small Brachiopoda fossile remains. In higher series of Lower Liassic according to facies changes, red, pink, brachipodal cherty limestone, red, pink massive crinoidal limestone and its crinoidal varieties of the "Hierlatz" type could be found.

Middle Liassic. It is known in the area of occurrence in two more important developments. A more frequent variety is the red, pink nodular or massive crinoidal limestone, the other the red, green-spotted cherty limestone. Sometimes small manganese grains appear in the crinoidal limestone, in several places manganese ore embeddings wedging out lenticularly are found.

Besides fully developed Lower and Middle Liassic, unconformity can also be experienced here. On the NE side of the "primary" oxide area karstic holes and fissures of Dachstein Liassic limestone are filled with Lower or Middle Liassic limestone and the Upper Liassic manganese ore series are deposited upon these unconformable border areas, too.

Upper Liassic. This series of the Jurassic is partly filled with manganese ore seams, deposited in haphazardly changing thickness upon the uneven surface of Lower and Middle Liassic formations. The overall thickness of the manganese ore series is 6—8 m, sometimes it even reaches 20 meters. Below in the ore series is found the manganese ore, deposited originally in oxide form from seawater accompanied by clay and clay marl as well, which is usually yellow, brownish below and darkgrey in the upper parts. The word "primary" which we have used and will use later indicates that manganese ore separated and sedimented originally in oxide form is in question, to be distinguished from the secondary manganese oxide ore, having been formed through oxidation of substances containing manganese carbonate. Upon the mentioned manganese series the dark grey radiolarian clay marl containing manganese carbonate bands, lenses, is sedimented. In the W and NE borders the ore series turns into manganiferous limestone and later barren limestone facies. Upon the manganese series already within Upper Liassic grey, red laminated limestone is settled.

Dogger-Malm. Lower Dogger is represented by red, nodular limestone with cherty lens with rich Ammonites fauna. Upper members of Dogger are represented by grey, greenish-grey and pink posidonian, laminated cherty limestone series. Upon this pink, white, grey cherty marl (radiolarite) representing Upper Dogger and Lower Malm (Bath-Callovian) is sedimented. The varied Eplény Jurassic sequence of series is closed by red, nodular ammonitic limestone.

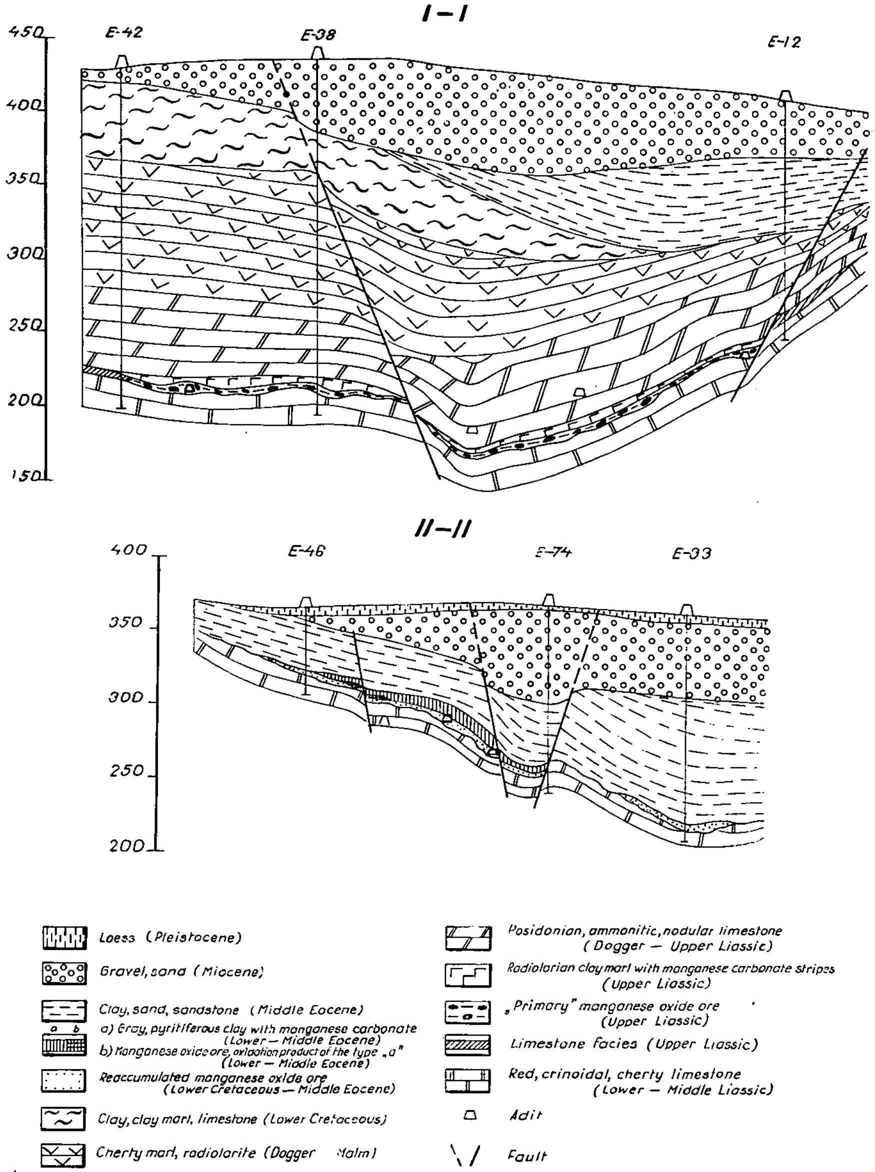

Fig. 2. Geologic sections of the investigated regions. *Fig. 1.*, I—I and II—II.

Cretaceous

In the area of occurrence Cretaceous is present as well in an insignificant thickness. The New-Kimmerian movements taking place upon the border Jurassic and Cretaceous had been quite decisive in forming the structure of the area although this period had played not such an important part in the history of development of the ore series in Eplény as in Úrkút. Cretaceous is represented by multicoloured clay, grey clay marl and limestone.

Eocene

Lower and *Middle Eocene* starts with transported and redeposited (reaccumulated) manganese series in the SE territory. Here one can find the mixed material of the manganese oxide deposit mentioned already in connection with Upper Liassic, together with the material of the overlying and the underlying rocks. Bauxitic clay

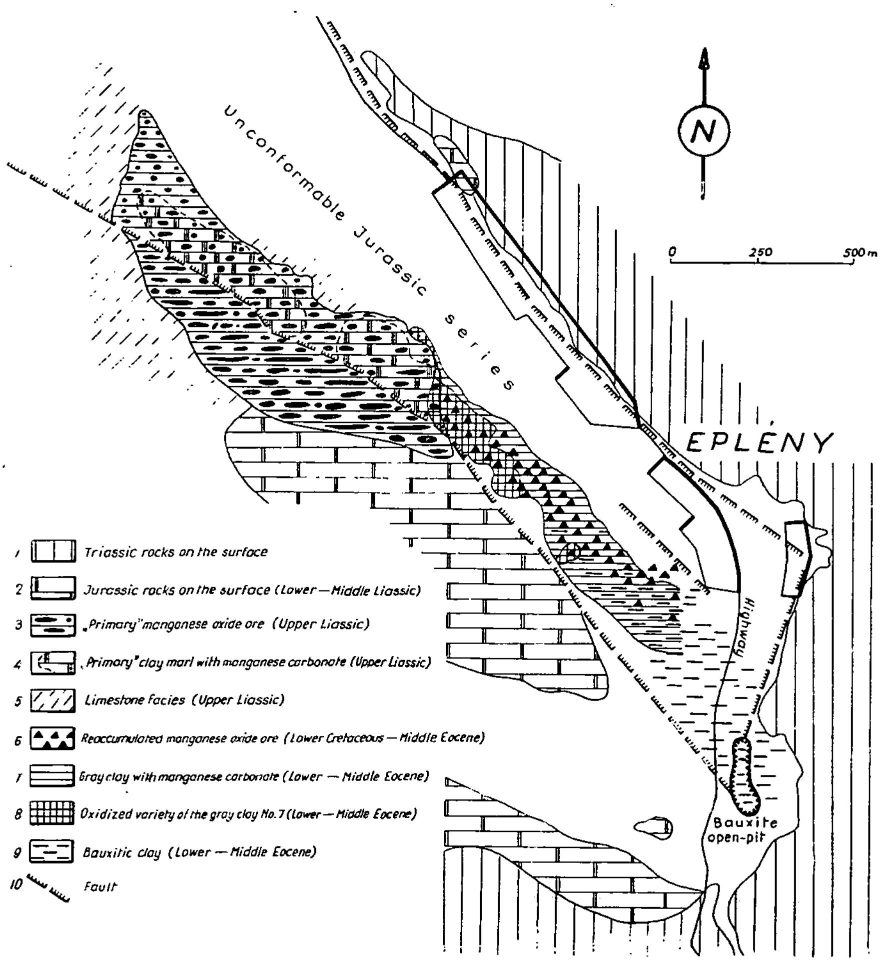

Fig. 3. A scheme of territorial distribution of Eplény manganese ore types

is settled upon this reaccumulated manganese series what formation is considerably thickened towards the abandoned surface bauxite mine. As a rule, grey, pyritiferous clay sometimes having considerable manganese carbonate content as well, is settled upon the bauxitic clay. On this substance carbonaceous clay, miliolinic sandstone and in slight thickness nummulitic limestone is settled and this ends the Eocene series.

Miocene

The Eplény manganese ore area is covered from the N by *Middle Miocene* gravel, sand and its maximum thickness reaches 70 m.

STRUCTURE OF THE AREA

The Eplény deposit is an organic part of the structural unit of the Bakony Mountains. The area is remarkably broken, considerably vertical and horizontal replacements could be observed. The structure of the area is characterized by two important, nearly parallel faults, of which the more decisive is the Géza-fault, cutting the area in the middle and the reaccumulated manganese series developed in the footwall of the fault. The other fault closes the Jurassic area from the NE and further on Triassic rocks can be found on the surface. The main direction of the faults is NW—SE. Perpendicular to this direction smaller breaks cut off the area.

NE from the manganese ore area the literature mentions an overlapped structure, namely in several borings Lower and Middle Liassic repeatedly is found. According to newer investigations in the unconformable areas holes, fissures, formed in Lower Liassic "Dachstein" limestone were filled up with Middle Liassic formations as indicated already.

The sequence of different series along I—I and II—II sections as given in *Fig. 1* is represented in *Fig. 2* and the areal distribution of manganese ore types is represented in *Fig. 3*.

DEVELOPMENT OF MANGANESE ORE BEDS

The area of the Eplény manganese ore deposit can be distributed into two groups according to their development.

1. Upper Liassic manganese ore beds of the western territory

a) "primary" manganese oxide ore
b) "primary" clay marl with manganese carbonate content

2. Lower Cretaceous — Middle Eocene and Lower Eocene — Middle Eocene manganese ore beds of the southeastern territory

a) reaccumulated (allochtonous) manganese oxide ore
b) grey, pyritiferous manganese carbonate-containing clay ("secondary")
c) manganese oxide ore ("secondary") formed by oxidation of grey, pyritiferous, manganese carbonate-containing clay.

1. Upper Liassic manganese ore beds of the western territory

a) Manganese oxide ore

In the area manganese had been settled already in oxide form from seawater upon the dissolved, corroded, uneven relief of the surface of the underlying rock (from this the name "primary"). The thickness of the manganese oxide series follows the unevenness of the underlying rock, reaching a maximum thickness of 5 m. Manganese ore is embedded in yellow, brown, black clays and dark—grey radiolarian clay marl. As to its structure two main types can be distinguished: there are known globular, reniform ore nodules while the other type is represented by layered, banked, loose manganese ores. The diameter of the ore nodules ranges from a few centimeters to 2,0 m, from outside they are covered with thin or thick ferruginous crust and in the inner parts we can find cherty — chalcedony parts. As a rule the manganese ore of loose structure is elongated and lenticular. Some transition forms between the two types could also be observed with a loose inner structure but showing globular structure from the outside and are delimited from the barren rock by a limonitic layer.

In the area of the south side of Géza-fault the loose, layered ore was more important while on the north side the globular, nodular ore type could be found more frequently. Passing from W towards NW within the oxide bed the ratio of the barren rock increases on behalf of the ore itself. The overall thickness of the series reducing it to the oxide ore is 1,6 m. Considering the total thickness of the oxide ore series the average quality of the manganese ore is the following:

Mn:	24,37%	Fe:	10,65%
P:	0,25%	SiO_2:	20,36%

Taking into consideration manganese ore quality without barren interbeddings, the manganese content of the ore as a rule changes between 30—36 per cent. Histograms constructed relying upon Fe, Mn, P, SiO_2 content determined from samples of the deposit *(Fig. 14—17)* very well show the frequency distribution of the mentioned components.

In connection with this deposit it is an interesting problem — to be clarified later on — that while oxide ore nodules are embedded into dark grey radiolarian clay at the same time differences in the conditions under which pyrite and the manganese oxides of higher valence were formed contradict the simultaneous formation. It can be supposed that formation of pyrite is the result of processes taking place in the clay after the separation of oxide ore nodules.

b) Clay marl with manganese carbonate content

Upwards in the "primary" oxide manganese ore series, oxide manganese ore nodules gradually disappear and dark grey radiolarian clay marl is found in the overlying, in the upper layer of which the grey, greenish-grey, yellowish-grey, finely layered manganese carbonate-containing clay marl usually appears lenticularly wedging-out. The manganiferous mineral is the extremely fine-grained rhodochrosite. In its development the manganese carbonate-containing clay marl is very similar to the Úrkút upper carbonate deposit, with a somewhat higher manganese content and a lower iron and phosphorus content. Its average quality:

Mn:	14,28%	Fe:	4,03%
P:	0,15%	SiO_2:	22,80%

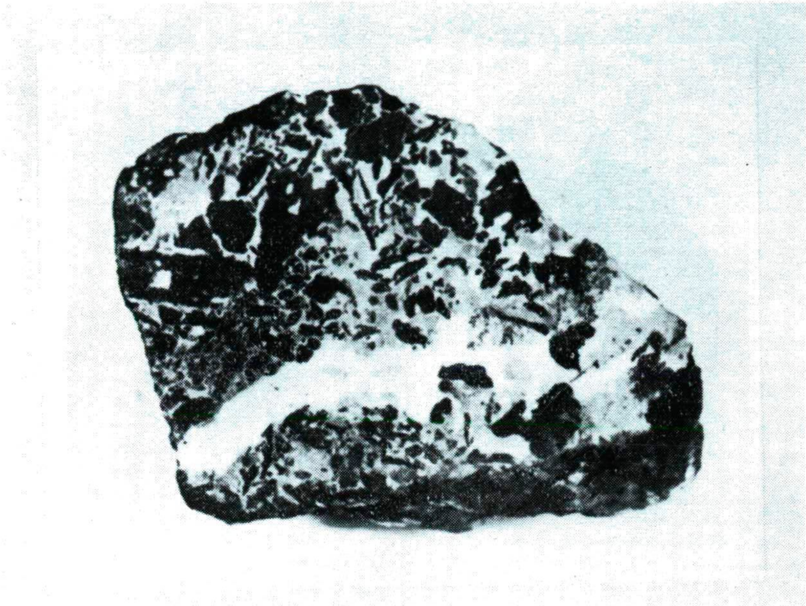

Fig. 4. Limestone showing brecciated structure with calcite and manganese ore as fissure and cavity fillings. About 1/3 natural size.

In the area between the W and SE mine fields the Upper Liassic formations and also a part of the manganese series became denuded. The manganese carbonate-containing clay marl which had got upon the surface likely had been oxidized, thus a band of layered oxide manganese ore of relatively low grade was formed. While in Úrkút it is just this ore type having great importance for the industry, in Eplény this type of ore has no significance from the point of view of practice. *Fig. 3* shows that manganese carbonate-containing clay marl is restricted to an insignificant area in Eplény, while in Úrkút the upper bed, corresponding to it, stretches over the main carbonate series. Manganese series is wedged-out towards the borders and as a rule turns into limestone facies. Towards the brims very often lonely manganiferous limestones appear under the oxide deposits with brecciated structure as can be seen in the photo reproduced in *Fig. 4*.

Geological sections of the Upper Liassic manganese series called "primary", at sampling places I and II in *Fig. 1,* are shown in *Fig. 5*.

2. Manganese ore beds of the southeastern territory

a) Lower Cretaceous — Middle Eocene reaccumulated allochtonous oxide manganese ore.

In the SE part of the Eplény area transported and redeposited manganese series is settled upon Lower and Upper Liassic karstic, eroded surface. Its material first of all had been given by the ore material of the previously already settled "primary" beds. In areas near the denuded Upper Liassic beds the ore debris is rather rough,

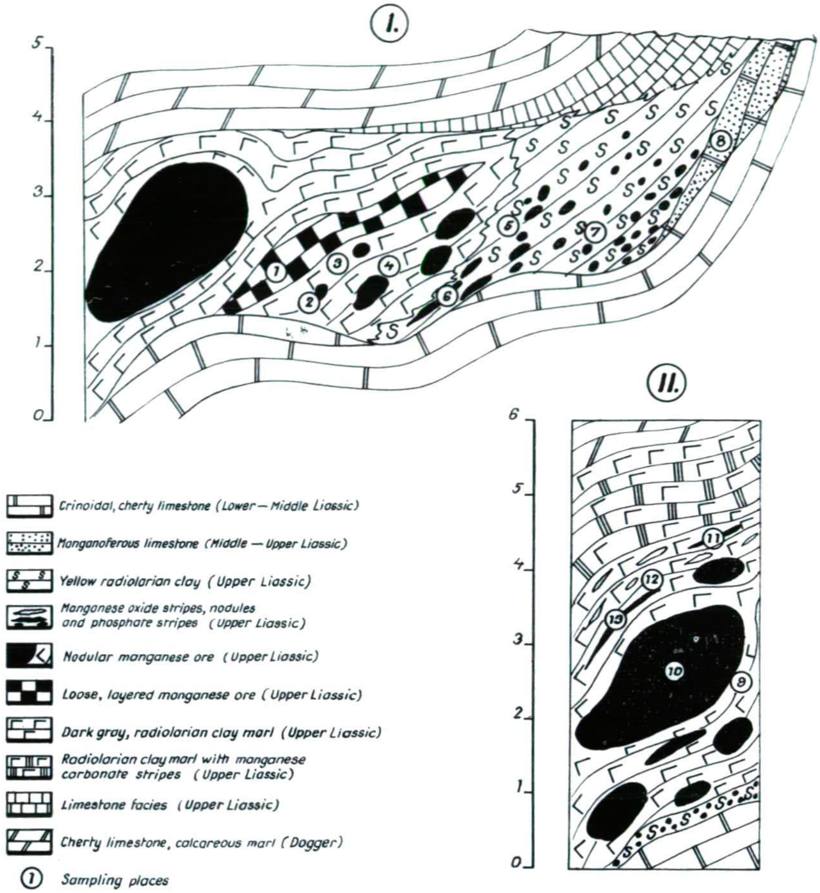

Fig. 5. Geologic sections of Upper Liassic manganese ore beds at sampling places I and II of *Fig. 1.*

with a diameter of 1—20 cm, moving on the grain size of the debris gets finer and other rock debris is mixed to it, mainly chert and limestone.

Generally the upper part of the bed consists of "muddy", very fine ore debris. The thickness of the reaccumulated ores waywardly changes — filling the karstic fissures of the bedrock it may even reach 5 meters thickness. The ore quality shows also a very strong change: for instance manganese content ranges from 17 to 31 per cent, while the phosphorus content is rather low (P: 0,02—0,07%).

In the upper part of the bed bauxitic material is settled upon the undulating surface.

b) Lower and Middle Eocene grey, pyritiferous manganese carbonate-containing clay

By the denuding of the first deposited Upper Liassic beds and later, through the aforementioned reaccumulation the Eocene transgression made possible to dissolve certain amount of manganese from the reaccumulated beds later again manganese carbonate-containing clay ("secondary") was accumulated by chemical processes. This material is first of all connected with the reaccumulated areas, but stretches over the "primary" beds as well and a variety of it with chert debris is known towards the brims. In part of the samples the manganese content ranges from 18 to 24 per cent what approximately corresponds to the quality of the Úrkút carbonate manganese ore, however, the Eplény manganese carbonate material has a considerably lower P and higher S content. Its average quality:

Mn:	15,05%	Fe:	9,30%
SiO_2:	15,57%	P:	0,08%
		S:	4,53%

c) Lower—Middle Eocene oxide manganese ore formed by oxidation of the grey, pyritiferous, manganese carbonate-containing clay

The grey, pyritiferous clay — containing manganese carbonate — is situated relatively near the surface on the Géza-fault side. Thus it became possible already after the sedimentation of the grey, pyritiferous clay, or somewhat later, manganese

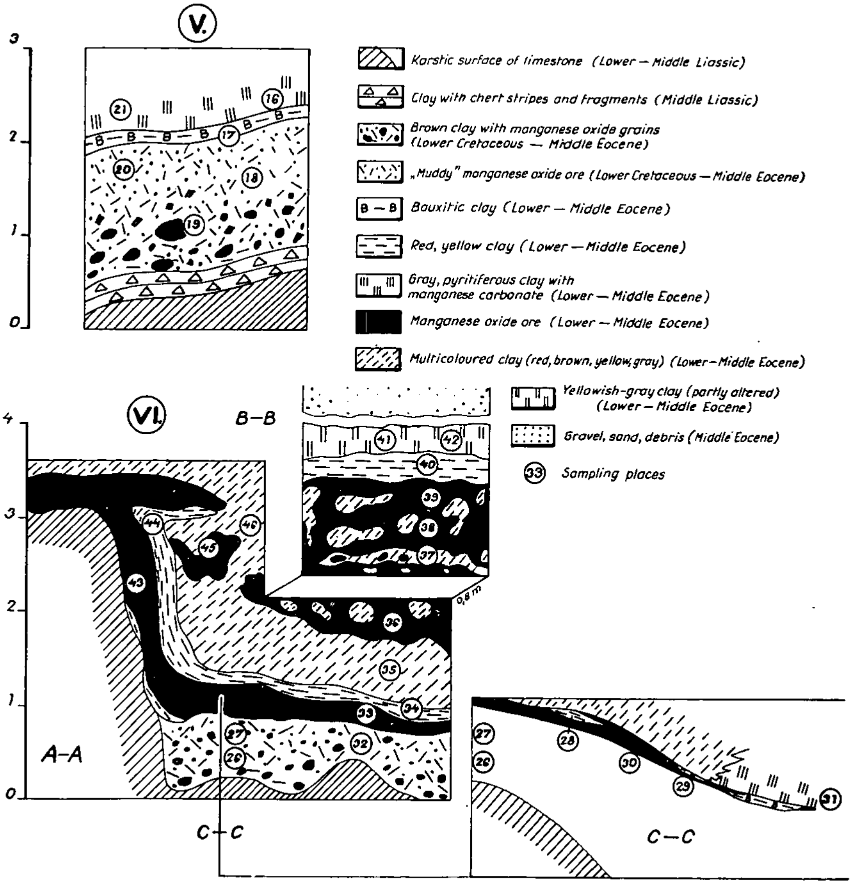

Fig. 6. Geologic sections of the SE territory at sampling places V and VI of *Fig. 1.*

carbonate being oxidized. At present near the shaft we know oxide ore ("secondary") formed due to the oxidation of pyritiferous manganese carbonate-containing clay, settled upon the reaccumulated oxide ore. This type of ore very frequently has a layered structure, its continuity is interrupted by irregular clayey interbeddings. The transformation of yellow, brown, red clays surrounding the oxide ore, into grey, pyritiferous clay can be followed everywhere. Similar changes are involved here as in the case of the Úrkút layered oxide ores and yellow radiolarian clay. The quality of the ore as a rule can be characterized as follows:

Mn:	25—28%	Fe:	8—10%
P:	0,1—0,2%	SiO_2:	18—20%

Geological section of the beds of the SE mine field at V and VI sampling places in *Fig. 1* is shown in *Fig. 6*.

POLLEN GRAINS OF THE CLAYS

Similar to investigations carried out with the Úrkút formations it appeared purposeful to orientate first of all as to organic remains, pollen grains, enclosed into the clayey members of the series. On our request, P. SIMONCSICS carried out pollen analytical study of several samples. Relying upon his information the following can be stated:

Dark grey radiolarian clay marl (No 3)* from the "primary" beds contains a number of organic fragments without structure. The number of spores—pollens, is rather few, their state is bad, they are creased, corroded, broken. The presence of the following types could be stated: *Leiotriletes sp., ? Cingulatisporites sp.* The determination of the following is uncertain: *Monosultices sp., Spheripollenites sp., Eucommiidites sp.,* while the presence of a "reticulate cista or alga" could be stated with certainty. All the above mentioned types could be found in Liassic, the type called "reticulate cista or alga" had already been detected by SIMONCSICS and KEDVES from the Úrkút deposit. The lack of the *Classopollis* genus is striking what may indicate that we are having sediments far from the shores.

In a grey, fine-layered material with considerable phosphorus content (No 11) also from the Upper Liassic, less organic remains were observed and the determination of the following genus (with very low occurrence) was also uncertain: *Monosulcites sp., Spheripollenites sp., Eucommiidites sp.* On the contrary, the determination of *Tittodiscus sp.* and *Classopollis torosus* is certain. The latter is mostly characteristic for the Liassic, later it hardly can be found. A single creased example — although the determination is uncertain — of *Multiporopollenites sp.,* was found. If the determination is correct, *Juglandaceae* pollen is possible, this, however, is impossible in Upper Liassic, it may perhaps occur only from Upper Cretaceous.

At the same time not only yellow Upper Liassic clay marl (No 5) from the "primary" series proved to be free of pollens, but the bauxitic clay settled upon the Lower — Middle Eocene "secondary" formation (No 17) and grey pyritiferous manganese carbonate — containing clay from the Middle Eocene "secondary" series as well (No 16 and 21), together with the grey pyritiferous clay (No 22) and yellow clay (No 23).

* Number in parenthesis here and in the following means sample numbers. Corresponding sampling places see in *Figs. 5, 6,* and short denotation of samples in Table 1.

TABLE I

Short denotation of the samples investigated

Serial number of the samples	Denotation of the samples	Sampling places: age and territory	Sampling places: shown by Fig.
1.	Loose, layered oxide ore	Upper Liassic, "primary" beds of the western mine territory	5. I.
2.	Oxide ore nodule with ferruginous crust		
3.	Dark grey, radiolarian clay marl		
4.	Oxide ore nodule with limonitic crust		
5.	Yellow, radiolarian clay marl		
6.	Layered oxide ore band from yellow radiolarian clay marl		
7*a*	Ore grains from yellow radiolarian clay marl		
7*b*	Oxide ore nodule from yellow radiolarian clay marl		
7*c*	Yellow radiolarian clay marl crust of the ore nodules		
8*a*	Manganese oxy-hydrate aggregate in white calcite		
8*b*	Brown, yellowish-brown, sandy, brecciated limestone with manganese ore grains and calcite veinlets		
9*a*	Compact, grey, crystalline manganese ore lens		5. II.
9*b*	Black, loose, earthy crust of the former		
10.	Loose, earthy oxide ore		
11.	Grey, layered phosphorus-bearing lens		
12.	Ferruginous lens with manganese ore bands		
13.	Sandy, grey, greenish-yellow stripped phosphorus-bearing lens		
18.	Rounded off manganese ore granules embedded in yellow, brown clay	Lower Cretaceous—Middle Eocene transported and redeposited, allochtonous beds of the southeastern mine territory	6. V.
19.	Dense, massive, black ore lens		
20.	Loose, earthy, "muddy" manganese oxide ore with yellow, brown clay		
26.	Rounded off manganese ore granules and dense, greyish ore fragments embedded in brown clay		6. VI.
27.	Brown manganiferous clay similar to the sample Nr. 26.		
32.	Manganese ore granules in brown clay		
17.	Red, bauxitic clay	Lower—Middle Eocene "secondary" beds of the southeastern mine territory	6. V.
16.	Grey, pyritiferous, manganese carbonate-containing clay		
21.	The same as sample Nr. 16.		
23.	Yellow clay from the transitional field		

TABLE 1—(continued)

Serial number of the samples	Denotation of the samples	Sampling places	
		age and territory	shown by Fig.
24.	Loose, oxidized, partly crystalline manganese ore from the transitional field		—
25.	Loose, oxidized manganese ore from the transitional field		—
28.	Red clay with manganese ore grains	Lower—Middle Eocene "secondary" beds of the southeastern mine territory	6. VI.
29.	The same as sample Nr 28		
30.	Clayey, oxidized, hard manganese ore		
31.	Grey, pyritiferous clay		
33.	The same as the sample Nr 30		
34.	Red clay with manganese ore grains, similar to the samples Nr 28 , 29		
36.	Clayey, earthy, in some places hard, black oxide ore		
38.	Hard, compact oxide ore pieces in red — white clay (a: red clay; b: white clay; c: ore)		
39.	Similar to the sample Nr 38		
40.	Pinkish clay with manganese ore grains		
41.	Light brown — greyish clay with manganese ore veinlets		
42.	Similar to the sample Nr 41		
43.	Black, clayey, layered oxide ore		
44.	Red-yellow clay		
45.	Clayey, oxide manganese ore		
46.	Manganese ore grains in yellowish-pinkish clay		

MINERAL COMPOSITION OF MANGANESE ORE SERIES

The study of mineral composition of manganese ores has been made difficult by the fact that most of the samples — with but a few exception — were earthy, porous, difficult to grind and polish and so it was impossible to prepare polished ore sections of corresponding quality. Thus in the study of mineral composition ore microscopy played by far not too important role and we could rely first of all upon results of differential thermal analysis and X-ray powder diffraction data, supplemented with chemical analysis of a number of samples as well.

D. T. A. curves taken with an ERDEY—PAULIK "Derivatograph" are shown in *Figs. 7, 8, 9* and *10.* Some of the X-ray powder diffraction patterns taken with CuKα radiation are given in *Figs. 11, 12* and *13.* Semiquantitative mineral composition of several samples and ore types, resp., is to be seen on the basis of X-ray powder diffraction data in Table 2.

Data of chemical analysis are summarized in Table 3., while the frequency distribution of some of the main components can be found in *Figs. 14—17.*

Places of sampling *(Nos 1—13)* from Upper Liassic "primary" beds in the W area are shown in sections I and II of *Fig. 5.* Places of sampling from allochtonous reaccumulated Lower Cretaceous—Middle Eocene beds in SE *(Nos 18—20, 26, 27,*

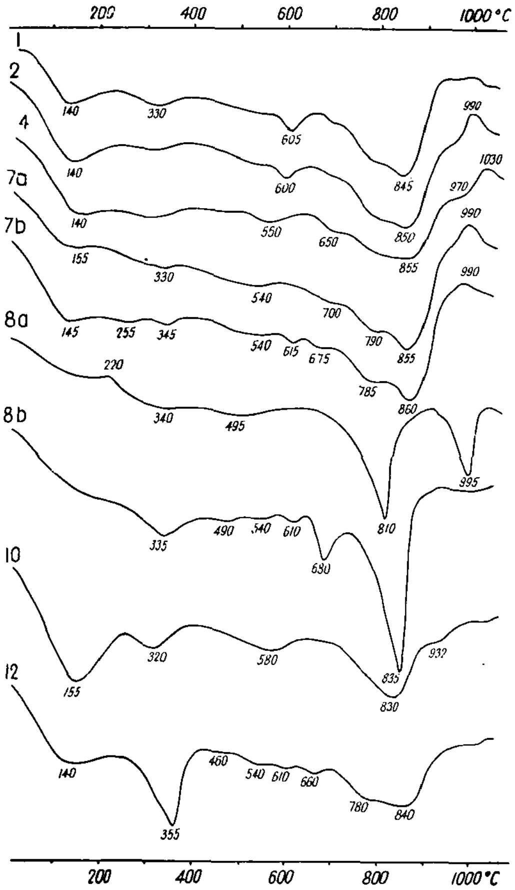

Fig. 7. D. T. A. curves of samples containing dominantly cryptomelane, collected from the Upper Liassic „primary" manganese beds (denotation of samples see in Table 1, their sampling places in *Fig. 5*).

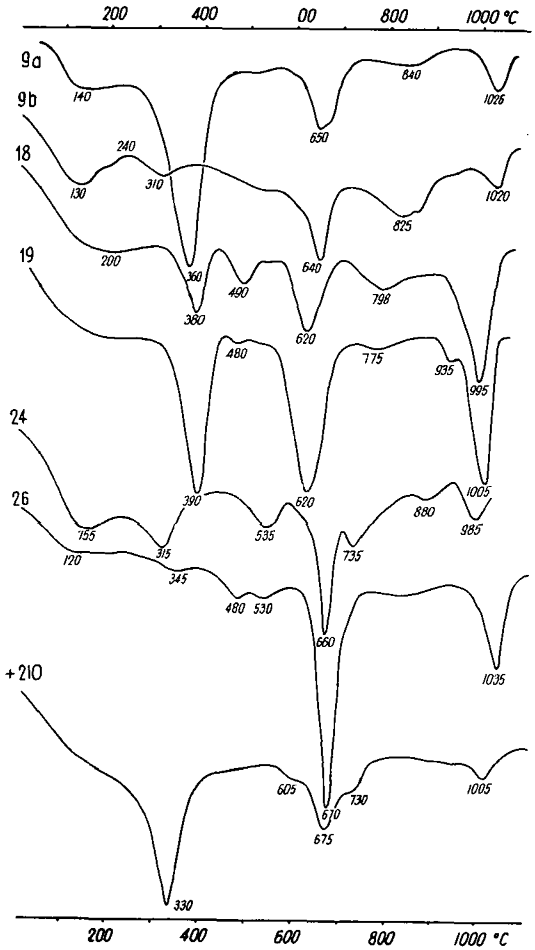

Fig. 8. D. T. A. curves of samples containing dominantly pyrolusite and manganite. Of the samples No *9a* originates from the "primary" bed, Nos *18*, *19* and *26* from the redeposited bed and No *24* from the "secondary" bed (denotation of samples see in Table 1, their sampling places in *Figs. 5, 6*).

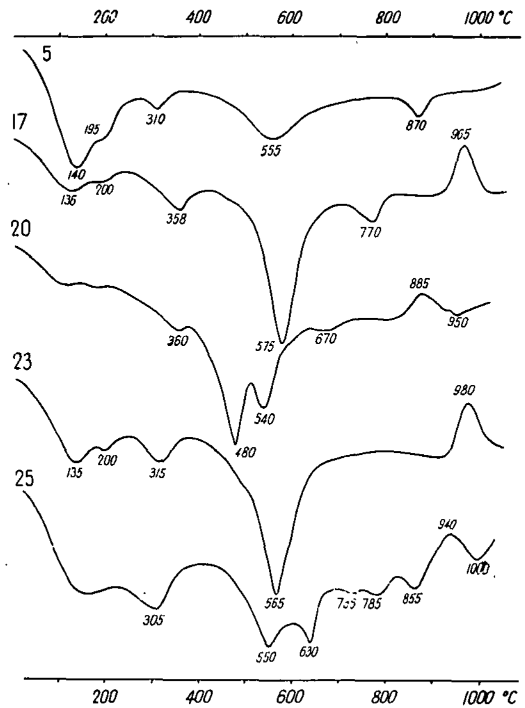

Fig. 9. D. T. A. curves of clayey samples. Of the samples No *5* taken from the "primary", Nos *17* and *20* from the redeposited and Nos *23* and *25* from the "secondary" beds. (Denotation of samples see in Table 1, sampling places in *Fig. 6.*)

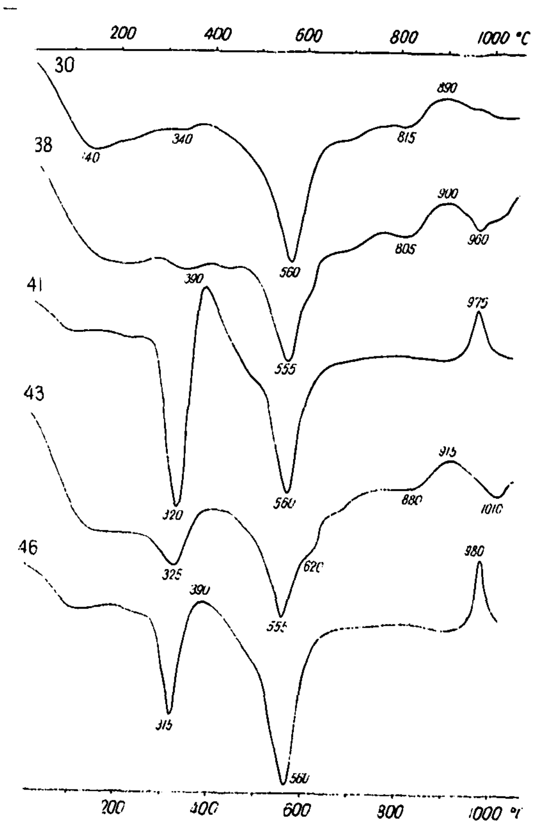

Fig. 10. D. T. A. curves of clayey samples from "secondary" beds. (Denotation see in Table 1, sampling places in *Fig. 6.*)

32) and from Lower—Middle Eocene "secondary" beds *(Nos 16, 21, 28—31, 33—46)* are shown in sections V and VI of *Fig. 6.*

On the basis of the D. T. A. investigations and X-ray powder diffraction data the composition of the ore can be outlined as follows.

Manganese minerals. In the western, Upper Liassic manganese ore series manganese had been deposited by all probability originally in the form of colloidal manganese oxy-hydrates. As it was already mentioned the manganese ore in this area is present on the one hand in layered, loose earthy form (e. g. samples *1, 6* and *10*) or sometimes in the form of lenses (samples *9* and *12*). However, smaller or larger globular, vesicular ore nodules usually with a ferruginous crust also occur (as samples *2, 4* and *7*). But independent of appearance, in the "primary manganese ore series cryptomelane appears to be the dominating manganese ore mineral as shown on the D. T. A. curves by the expressed endothermic peak between 780—860 °C, as shown in *Fig. 7.*

It could be observed even in polished ore sections of low quality that the matrix of both the loose, earthy, layered type and of the globular, vesicular harder ore nodules is cryptocrystalline cryptomelane containing more or less limonite. The colloform structure suggests a separation from the colloidal state. Alongside the layers or throughout the ore nodule veinlets harder than the matrix are found. The material of these veinlets sometimes is pyrolusite, cryptomelane and sometimes manganite. Sometimes manganite needles or fine fibrous radiated aggregates of pyrolusite and of other MnO_2 modifications, respectively, can be observed.

Besides cryptomelane, pyrolusite, psilomelane and manganite the presence of todorokite and lithioforite is also obvious what has not been detected up to this time either in the Úrkút or the Eplény manganese ores. GY. BÁRDOSSY [1968] examining manganese ore spots in bauxite found near Eplény first pointed out in them the presence of lithioforite and starting from it the identification of lithioforite was successful in ores of the presently studied area, first of all by X-ray powder diffraction patterns, as shown in *Fig. 11.*

X-ray investigations of the manganese minerals of the seams pointed out also the presence — if sometimes in a rather slight amount — of nsutite, groutite, manganoan calcite and rhodochrosite as well.

The comparison of the Upper Liassic manganese seams in the W and manganese ores found in the redeposited and "secondary" beds of the SE territory suggests that there are certain interdependences between mineral composition and areal distribution. Namely while in the Upper Liassic so called "primary", series, relying upon the results of investigations, cryptomelane appeared to be the most frequent and besides it other manganese oxide minerals play only secondary role, among manganese ore minerals of the redeposited Lower Cretaceous — Middle Eocene series the amount of manganite and pyrolusite appears to grow, then in the "secondary" Lower—Middle Eocene beds again cryptomelane and psilomelane are dominating, pyrolusite and manganite being only subordinate. Lithioforite, todorokite, nsutite can also be found in seams of the southeastern field and in sample *No 24* the X-ray examinations detected the presence of much lithioforite as well, or for example in sample *No 25* the amount of nsutite is also considerable as it can be seen in the X-ray diffraction pattern shown in *Fig. 11.* Manganoan calcite, rhodochrosite are also present naturally in a greater quantity in manganese carbonate-containing clays.

Clay and bauxitic minerals. Examination of clay minerals showed that here as well certain areal distribution can be established. As it could be unequivocally stated from the D. T. A. curves of the corresponding samples and from data summarized in Table 2., among clay minerals of the Upper Liassic "primary" territory montmorillonite and illite are the characteristic ones, kaolinite plays only a subordinate role. On the contrary, in the southeastern Lower Cretaceous—Middle Eocene redeposited and Lower—Middle Eocene "secondary" beds the dominating clay minerals are kaolinite + sudoite (the amount of the two has not been determined separately) and alongside the *b* axis disordered kaolinite "*b*" (fireclay).

To distinguish between kaolinite + sudoite and kaolinite "*b*" (fireclay) was possible not only on the basis of X-ray powder diffraction data but also from the D. T. A. curves. In samples in which kaolinite and sudoite were involved, the exothermic peak between 965—980 °C indicating the formation of mullite, is sharp, while in D. T. A. curves of samples containing kaolinite "*b*", this peak temperature is lower (880—940 °C) and is not so expressed as in the case of kaolinite. All this can very well be seen when comparing D. T. A. curves of samples *No 30, 38* and *43* with those of *Nos 41* and *46*.

This distribution likely can be connected with differences between the conditions of separation and sedimentation. According to the suppositions manganese ore from the Upper Liassic was separated from seawater in colloidal state in form of oxide-hydrates. Thus this condition — first of all the pH — promoted the formation of montmorillonite while the conditions of sedimentation and later of alteration in the SE "secondary" Lower and Middle Eocene favoured the formation of kaolinite.

As it has already been mentioned previously between the transported and redeposited seams from the Lower Cretaceous—Middle Eocene and the "secondary" series of the Lower—Middle Eocene there is bauxitic clay which markedly thickens towards the abandoned bauxite open-pit. *(Fig. 3.)* In connection with this it can also be stated that in samples where the dominating clay mineral is kaolinite + sudoite and "fireclay" the X-ray examinations showed also the presence of gibbsite, boehmite, diaspore, the amount of which sometimes may be quite remarkable (samples *38 a-b; 41; 46*).

Other minerals. Besides clay and bauxite minerals the presence of goethite, pyrite, hematite, calcite, quartz and anatase in various amount can be detected. What about the limonite incrustation of the nodular ores, from the temperature values of endothermic peaks between 300 and 400 °C in the D. T. A. curves it can be supposed that besides crystalline goethite, pointed out by X-ray investigations and observed under ore microscope, also the presence of lepidocrocite or even more of amorphous limonite can be considered. This is even more suggested by the lower peak temperature (305—330 °C) than in the case of goethite meaning dehydration of FeOOH in some of the samples and then after this peak by the appearing of an exothermic peak, what can be interpreted by recrystallisation.

Mention also must be made of the phosphorus-bearing grey, greenish — yellow bands in the Upper Liassic series (samples *No 11* and *12*) between the radiolarian clay marl with manganese carbonate content and the dark-grey radiolarian clay marl enclosing the ore seams. Examining the phosphorus-bearing mineral of these bands earlier the presence of fluorapatite had been supposed. It was unequivocally stated by GY. GRASSELLY [1968] by detailed chemical, X-ray powder diffraction data and infra-

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Abbreviations used in the X-ray powder diffraction patterns of *Figs. 11, 12* and *13* are as follows: An: anatase; Boe: boehmite; Cal: calcite; Cr: cryptomelane; Gi: gibbsite; Goe: goethite; Gr: groutite; Gy: gypsum; Ill: illite; Ka: kaolinite; Ka „b": kaolinite "b" (fireclay); Li: lithiophorite; Mn: manganite; Mn—Cal: manganoan calcite; Ns: nsutite; Py: pyrite; Pyrol: pyrolusite; Q: quartz; Rhod: rhodochrosite; Su: sudoite; Tod: todorokite.

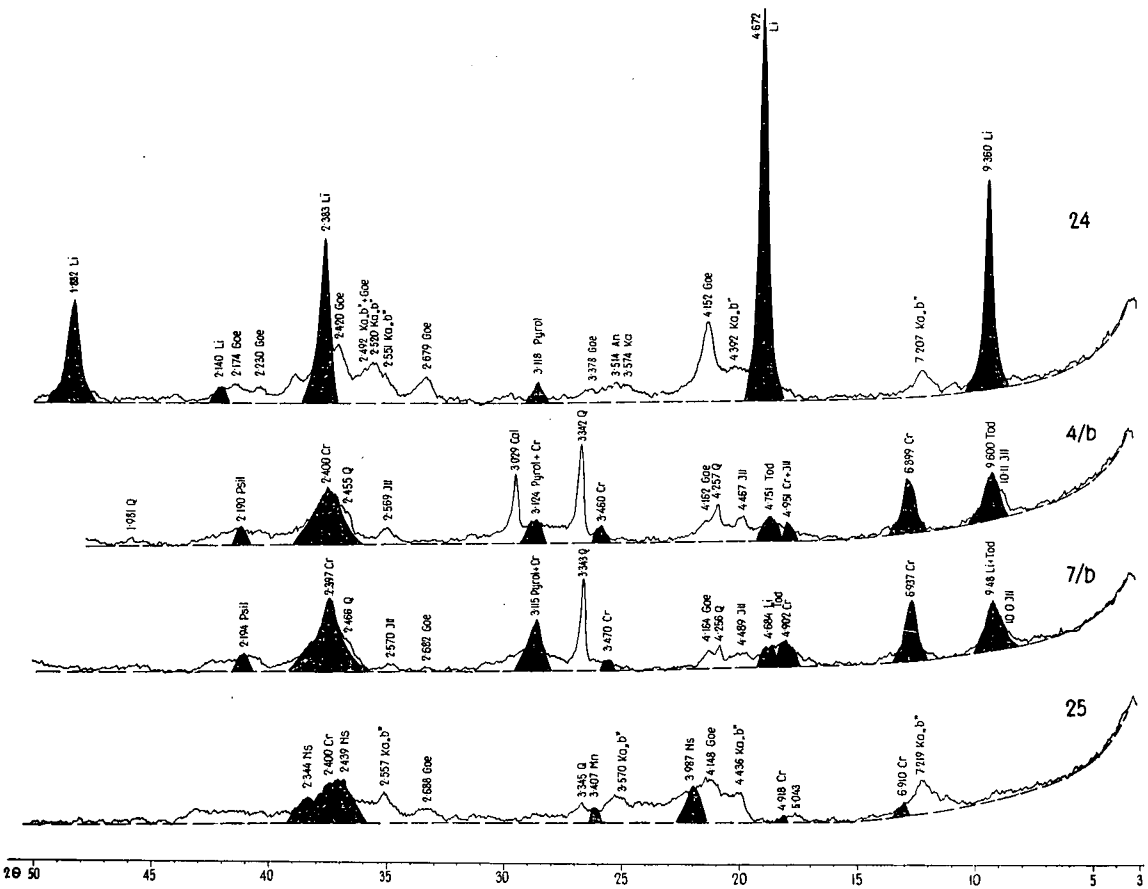

Fig. 11. X-ray powder diffraction patterns of samples containing cryptomelane, lithioforite, todorokite, nsutite.

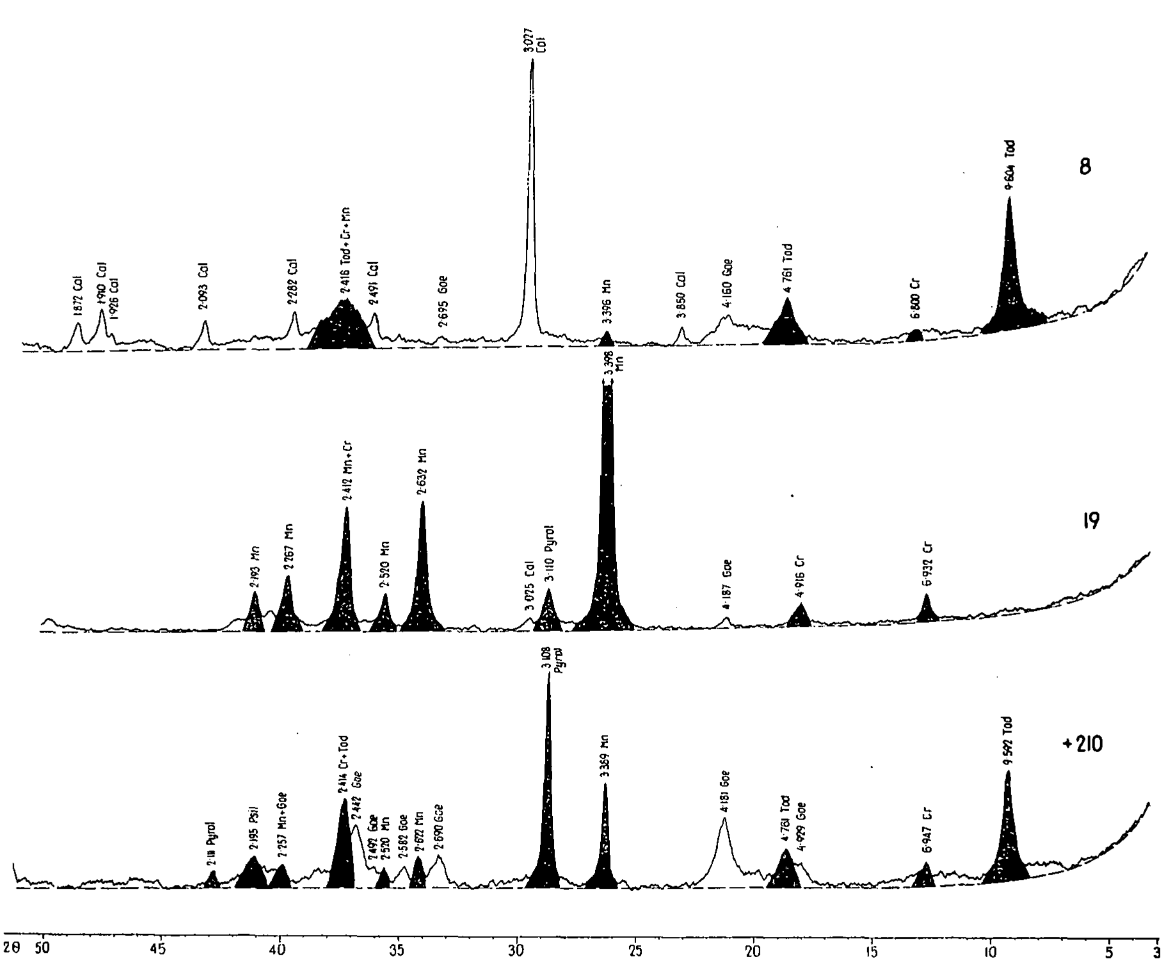

Fig. 12. *X*-ray powder diffraction patterns of samples containing pyrolusite, manganite, cryptomelane and todorokite.

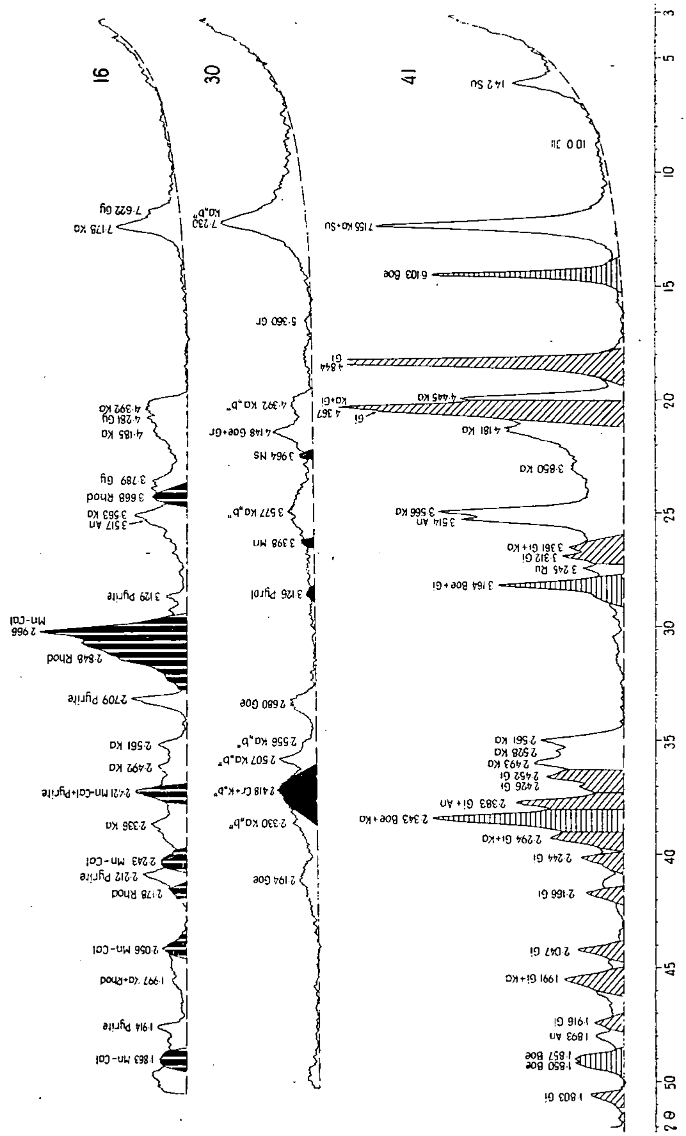

Fig. 13. X-ray powder diffraction patterns of samples containing kaolinite + sudoite, kaolinite "*b*" (fireclay), gibbsite, boehmite, manganoan calcite.

TABLE 2

Semiquantitative estimation of the mineral composition of some samples on the basis of X-ray powder diffraction patterns

Serial number of the samples	Estimated amount of the components						
	>80%	80—60%	60—40%	40—20%	20—8%	8—2%	<2%
	I. Upper Liassic "primary" beds of the W ore field						
3.			montmorillonite	illite	quartz, pyrite	manganoan calcite, feldspar	
4*b*[1]			cryptomelane	todorokite	illite	psilomelane, pyrolusite, goethite, quartz, calcite	
5.				montmorillonite, illite	quartz, goethite	kaolinite	manganoan calcite, rhodochrosite, feldspar
6.				cryptomelane, pyrolusite, lithioforite + todorokite (together)	illite	quartz	
7*b*[1]			cryptomelane	lithioforite + todorokite (together)	pyrolusite	psilomelane, goethite, quartz	
7*c*			montmorillonite	quartz	illite goethite	manganoan calcite, feldspar	
8*b*[2]				todorokite, calcite		cryptomelane, goethite	manganite

9*a*			pyrolusite	goethite	cryptomelane	psilomelane, groutite, lithioforite	quartz
10.			cryptomelane		psilomelane, illite, manganite	pyrolusite, lithioforite, goethite, quartz, calcite	
II. Lower Cretaceous- Middle Eocene reaccumulated allochtonous beds of the SE ore field							
19.[2]		manganite			cryptomelane, pyrolusite	goethite, calcite	
20.			pyrolusite	psilomelane	kaolinite, illite	goethite, quartz	
III. Lower Cretaceous- Middle Eocene „secondary" beds of the SE ore field							
17. *a*		kaolinite			hematite, goethite	illite, anatase, calcite	
17. *b*	kaolinite+ sudoite					calcite, goethite, anatase	
16.[3]				kaolinite, manganoan calcite	rhodochrosite	pyrite, gypsum	anatase
24.[1]			lithioforite		kaolinite "b" (fireclay) goethite, pyrolusite		

[1] X-ray diffraction pattern of these samples see in *Fig. 11*.
[2] X-ray diffraction pattern of these samples see in *Fig. 12*.
[3] X-ray diffraction pattern of these samples see in *Fig. 13*.

TABLE 2 — (continued)

Serial number of the samples	Estimated amount of the components: >80%	80—60%	60—40%	40—20%	20—8%	8—2%	<2%
	III. Lower -Middle Eocene "secondary" beds of the SE ore field						
25[1]				cryptomelane, nsutite, kaolinite "b" (fireclay)	goethite	manganite, quartz	
30[3]			kaolinite "b" (fireclay)	cryptomelane		nsutite, pyrolusite, goethite, groutite, manganite	
38. *a*		kaolinite + sudoite			goethite, hematite	gibbsite, boehmite, anatase	diaspore
38. *b*		kaolinite + sudoite			gibbsite	goethite, anatase	
38. *c*				kaolinite "b" (fireclay) cryptomelane	todorokite	pyrolusite, gibbsite	
41[3]		kaolinite + sudoite		gibbsite	boehmite	illite, anatase	
43.				kaolinite "b" (fireclay) cryptomelane, psilomelane	goethite	pyrolusite, manganite, gibbsite	boehmite
46.		kaolinite + sudoite			gibbsite	goethite, anatase	

[1] X-ray diffraction pattern of these samples see in *Fig. 11.*
[2] X-ray diffraction pattern of these samples see in *Fig. 12.*
[3] X-ray diffraction pattern of these samples see in *Fig. 13.*

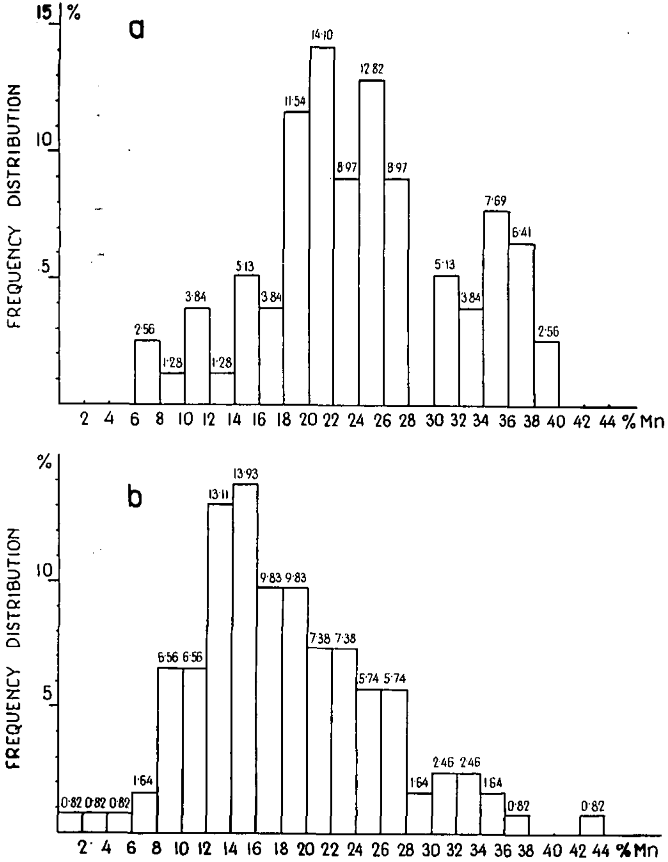

Fig. 14. Mn frequency distribution in the "primary" *(a)* and in the redeposited *(b)* region.

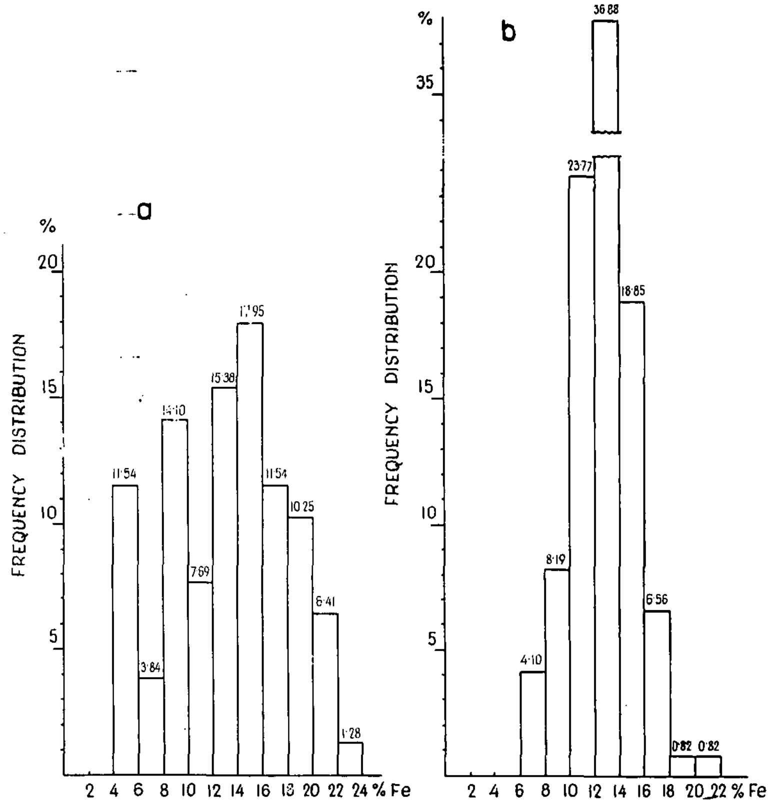

Fig. 15. Fe frequency distribution in the "primary" *(a)* and in the redeposited *(b)* region.

TABLE 3

Fe_2O_3, *MnO, active* O *and alkali content of some samples*

Serial number of samples	Fe_2O_3 %	MnO %	O %	Na_2O %	K_2O %	MnO_x x=
1.	3,99	58,42	10,66	0,35	3,5	1,808
5.+	10,25	not determ.		0,19	3,47	—
6.	5,58	53,58	9,83	0,3	2,87	1,812
7*a*	3,79	59,84	11,17	0,35	2,85	1,827
7*b*	3,82	63,06	11,45	0,35	3,52	1,804
8*a*	—	58,61	11,69	0,31	0,67	1,884
8*b*	6,91	41,96	8,44	0,36	0,94	1,891
9*a*	50,75	22,40	4,15	not determ.		1,821
9*b*	9,97	42,15	8,61	not determ.		1,905
10.	12,17	33,89	7,12	0,32	3,30	1,931
12.	27,60	41,43	6,52	0,15	2,10	1,920
15.+	38,05	8,07	1,46	—	1,67	1,801
18.	2,59	66,16	10,99	0,1	2,04	1,736
19.	2,47	71,65	10,22	0,1	0,85	1,632
20.	17,14	24,14	3,96	0,2	0,3	1,727
24.	19,59	49,70	10,23	0,12	0,4	1,912
25.	16,20	33,24	6,84	0,18	0,8	1,912
30.+	9,54	28,79	5,55	0,25	0,54	1,854
38.	10,05	29,43	6,07	0,69	0,3	1,914
41.+	2,67	—	—	0,05	0,2	—
43.	24,61	21,75	4,68	0,5	0,75	1,954

Remarks: Data of samples denoted with+ refer to average samples, the other data to selected material.

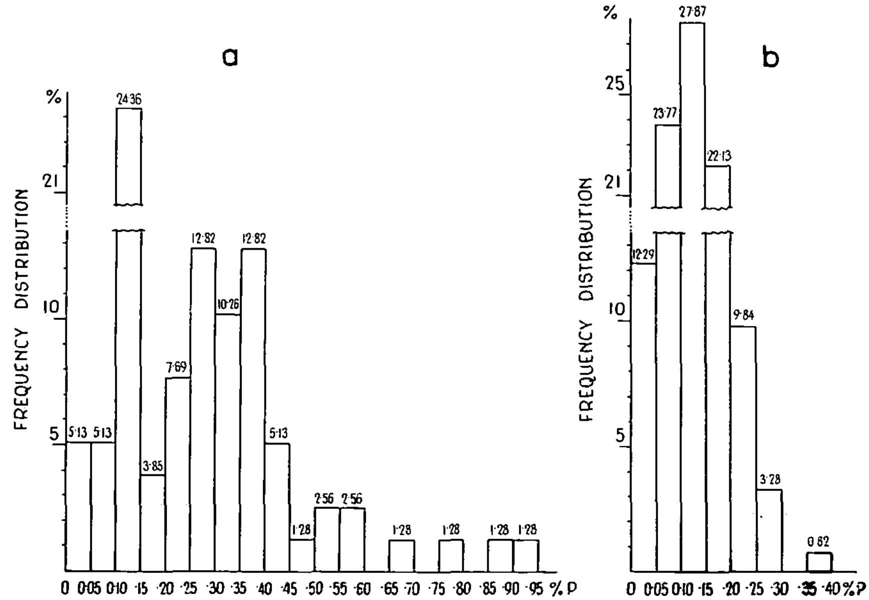

Fig. 16. P frequency distribution in the "primary" *(a)* and in the redeposited *(b)* region.

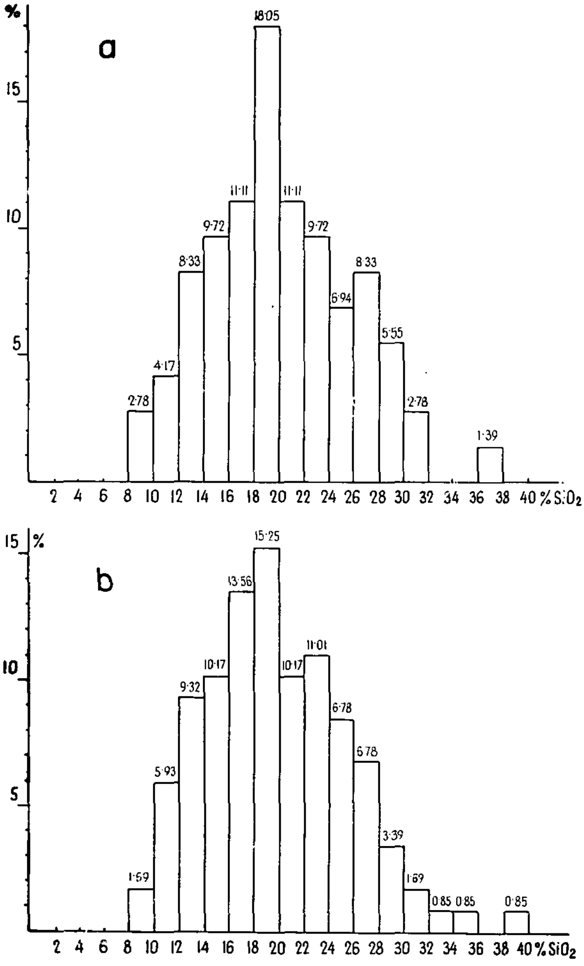

Fig. 17. SiO_2 frequency distribution in the "primary" *(a)* and in the redeposited *(b)* region.

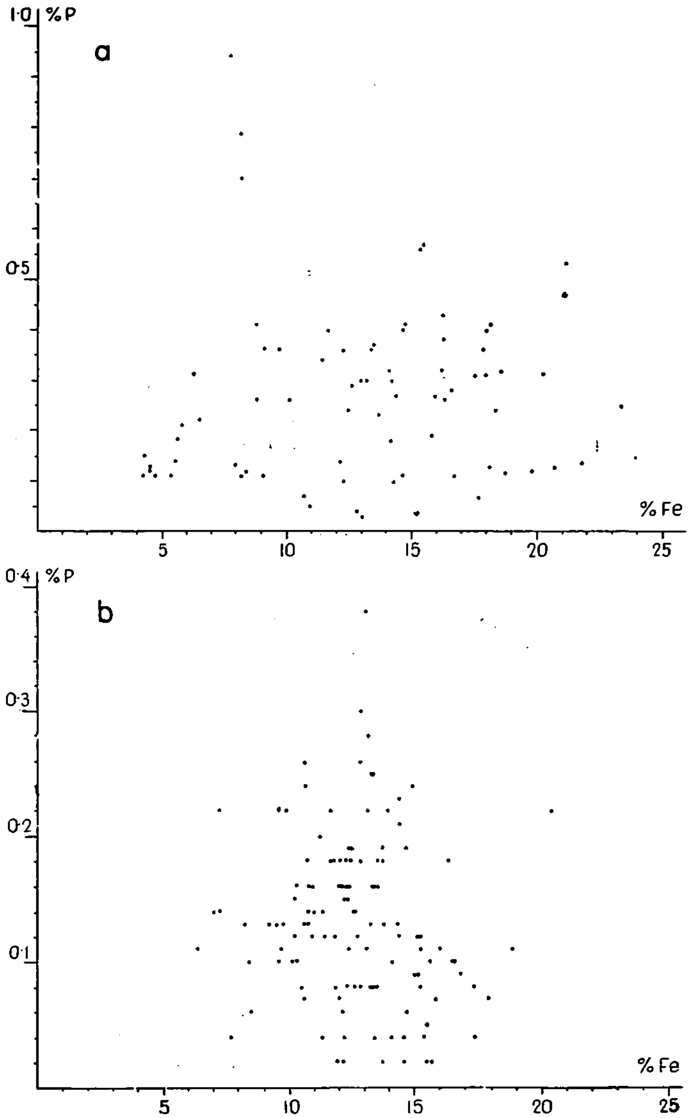

Fig. 18. Correlation between Fe and P content in the "primary" *(a)* and in the redeposited *(b)* region.

red spectroscopy that in the Úrkút and Eplény fields the phosphate-mineral is hydroxy-carbonate-apatite, with a low F-content, which is very near dahllite.

Chemical analysis first of all aimed to establish whether there is any difference in the distribution of Fe, Mn, P, SiO_2 between the "primary" area and the redeposited beds. On determining the above mentioned components in 78 samples taken from the "primary" beds and in 122 samples of the redeposited area, the frequency distribution is shown by the histograms represented in *Figs. 14—17.*

On comparing histograms of these two areas it can be stated that those of the primary area, either in the case of Mn, Fe or P have several maxima, while, on the contrary, in the redeposited area on the one hand the chief maximum appears at a lower percentage and the histogram will be more balanced, near unimodal on the other. In the course of the transportation and reaccumulation of the material of the „primary" beds — as it has already been mentioned above — partly manganese could be dissolved: the overall manganese content of the reaccumulated manganese beds is somewhat lower than that of the "primary" beds. At the same time the process of transportation, redeposition, dissolution and precipitation connected with these, may result in a more even distribution of the components. The only exception is SiO_2, in the distribution of which no marked difference could be established relying upon the histograms.

A very interesting picture can be obtained concerning the correlation between Fe and P content in the two areas *(Fig. 18)*. On the basis of 78 samples collected from the "primary" territory certain positive correlation seems to exist and although the points are rather scattered, as a rule the P content increases with the Fe content. Similar interdependence was pointed out by GY. GRASSELLY and J. CSEH NÉMETH [1961] in the manganese ore beds of the Úrkút slope, as was also stated by ASOKA MOOKHERJEE in the Indian gondites [1961]. This latter author pointed out that this connection can be interpreted by the chemisorption of the PO_4 anion by colloidal, positively charged $Fe(OH)_3$. In contrary to this Fe per cent *versus* P per cent does not show such a connection in the reaccumulated area. It seems probable that a correlation between iron and phosphorus content can be experienced only in territories where iron precipitated mostly in colloidal from and PO_4 also reached in the solution the required concentration and so through chemisorption there was a possibility for PO_4 being sorbed. In areas, however, as e.g. the redeposited beds of the SE field in Eplény, where rough or fine debris of the "primary" manganese ore seams and that of the overlying and underlying rocks were already in solid form and redeposited again and where colloidal iron hydroxide did not play such a part as in the "primary" seams, the adsorption of PO_4 did not take place, even more when both in the transporting and depositing medium PO_4 concentration decreased below the value necessary for the flocculation of colloidal $Fe(OH)_3$.

Detailed analysis has been made from several samples first of all to determine the total amount of manganese oxides, the manganese — iron ratio and the overall grade of oxidation of manganese ores coming from different series. Data are summarized in Table 3. The grade of oxidation at least suggests that in oxide ores mostly manganese oxides — oxy-hydrates of higher valence — are found and the Mn:O ratio is under the value $MnO_{1,8}$ only in the case of a few samples, first of all where considerable manganite content is involved.

Summarizing, the formation of the Eplény manganese ores can be outlined as follows:

Chemical separation and sedimentation from seawater

Upper Liassic manganese formation
"primary"

MANGANESE OXIDE ORE

Denudation of overlying rocks

Transportation and redeposition

Upper Cretaceous
Middle Eocene

CLAY MARL
with
MANGANESE CARBONATE

Oxidation on or beneath the surface

LAYERED MANGANESE
OXIDE ORE
autochtonous

Transportation and redeposition

REDEPOSITED MANGANESE OXIDE ORE
allochtonous

Partial dissolution of the ores during transportation and redeposition

Under favourable conditions (Eh, pH) separation of manganese carbonate from the solution

GREY, PYRITIFEROUS CLAY
with
MANGANESE CARBONATE

Lower — Middle Eocene

Oxidation on or beneath the surface

MANGANESE OXIDE ORE
"secondary"

ACKNOWLEDGMENTS

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TECTONICS OF THE NORTHWESTERN SLOPE OF THE MÁTRA MOUNTAINS

J. Mezősi

INTRODUCTION

The investigation area lies in the northwestern part of the Mátra Mountains mass, as shown on the attached map-scheme *(Fig. 1)*. In the west, the morphological patterns and the disposition of the formations reveal that both the Várhegy of Hasznos and the volcanic mass in the vicinity of Tar are in a markedly down-dropped position with regard to the Mátra Mountains mass. The fracture system of the Zagyva Graben is also indicative of tectonic movements. Consisting mainly of andesites and their tuffs and subordinately of dacitic tuffs, the Helvetian to Lower Tortonian volcanic complex is presently about 300 m thick. It rests on the schlier formation overlying the Helvetian lignite formation. These formations are locally covered by a thick talus mantle concealing both the majority of the andesite dykes of the Mátra Mountains sedimentary foreland and the tectonic lines intersecting the formations.

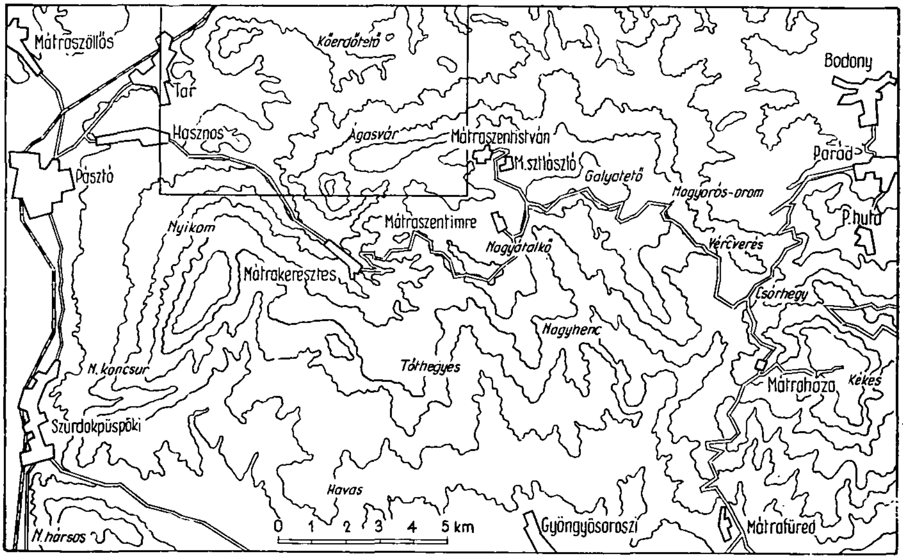

Fig. 1. Geologic map of the area investigated.

As shown by drilling in the broader environs of the Mátra Mountains (Sóshartyán, Szécsény) and by the xenoliths recovered from the deeper portions of the Mátra Mountains andesites, the basement of the northwestern sector of the mountains seems to be constituted by crystalline schists [K. BALOGH, 1966. and BALOGH—KŐRÖSSY, 1968.]. The eroded surface of these schists, has been overlain by Paleogene sediments.

The oldest formation uncovered by drilling near the Mátra Mountains northwestern border is the clay — clay-marl — sandstone sequence of the Rupelian stage which was hit at 401.50 m depth during the drilling of borehole Nagybátony-I, but which was not yet cut through at 1537 m, where drilling was stopped. It is overlain first by an Upper Oligocene sequence, then by 250 m of Burdigalian deposits. The lower part of the last-mentioned formation is represented by marine sediments with larger pectinids, the upper part, in turn, by 16 to 60 m of so-called Lower Dacitic Tuff of peculiar white colour which used to be referred to as "Lower Rhyolite Tuff". Exploratory drilling was stopped at the point where the dacitic tuff underlying the Helvetian lignite formation was reached, so that little is known about their actual thickness and facies.

After the Late Burdigalian continental phase the Helvetian epoch began by a slow ingression of the sea. The slow subsidence of the region is indicated, on the one hand, by the clay mineralization of the upper portion of the dacitic tuffs (a phenomenon suggestive of inundation!), on the other hand, by the deposition of lignite seams.

Whereas in the vicinity of Mátranovák and Homokterenye the lignite formation consists of three seams, at the Mátra Mountains northern foot it includes only two of them. Seam II is of deep-bog origin, including a barren intercalation of 0.7 to 0.8 m thickness. The higher-seated Seam I is of shallow-bog origin. The seams have dip angles of $165°-185°/6°-12°$.

After the deposition of the lignite formation the rate of subsidence was accelerated, which gave rise to gradual pinching out in southern direction of the lignite. The lignite formation is overlain, after a thin intercalation of Chlamys sands (which may locally lack), by marly, micaceous siltstones (schlier) which usually contain a poor fauna. In Csutaj pit and Szalajka brook the author of the present paper collected the following fossils which were determined by M. BOHN—HAVAS [MEZŐSI, 1966]: gastropods — *Turritella beniosti* COSSMANN et PAYROT, *Turritella subangulata* BR., *Ringicula (Ringiculella) auriculata buccinea* BR., *Rissoa sp.*, *Neritina sp.*, *Turritella sp.*, *Polynices sp.*, *Columbella sp.*, *Cantharus sp.*, *Drillia sp.*, *Natica sp.*, *Clavatula sp.*, bivalves — *Venus (Clausinella) basteroti* DESH., *Solenocurtus candidus* REN., *Pinna sp.*, *Venus sp.*, *Tellina sp.,;* foraminifers — *Robulus sp.*, *Nonion sp.*, *Nodosaria sp.,;* fragments of Ostracoda and Echinus. From another locality of the Szalajka Valley, I. CSEPREGHY—MEZNERICS [1954] quoted the following mollusc species: *Protoma cathedralis paucicincta* SACCO, *Architectonica simplex* BRONN., *Nassa (Usita) restituta hörnesi* MAYER, *Nassa (Caesia) cf. inconstans* HOERNES et AUINGER, *Ancilla (Baryspira) glandiformis* LAMARCK, *Conus (Conospira) dujardini* PHIL., *Ringicula (Ringiculella) auriculata buccinea* BROCCHI, *Leda (Lembulus) fragilis* CHEM., *Angulus (Oudardia) compressa* BROCCHI. In the upper reaches of Szalajka brook SCHRÉTER [1940] found *Brissopsis sp.* specimens, north of Kőerdő Hill he collected *Ringicardium danubianum* MAYER and MACOMA *elliptica* BROCCHI var. *ottnangensis* R. HOERN.

In the geological survey borehole Hasznos-I, in Helvetian schlier, the following forms were found and determined by M. MUCSI: *Stenothyra sp.*, *Potamides sp.*, *Venus sp.*, *Tellina sp.*, *Arca sp.*, *Cardium sp.*, and *Pinna sp.*

Ingression was replaced by regression, coupled this time with andesitic volcanism, as early as the second half of the Helvetian. The first layers of the about 60- to 100-m-thick agglomerated andesite tuffs and andesites are still intermingled with marine sands and clays; these tuff layers are stratified, their material is graded. The best exposure is in the Csevice Valley near Tar and in the vicinity of Tyukod Hill. The tuffs are also represented in the core of the survey borehole Hasznos-l (southern slope of Hegyes Hill). At the time of lava effusion that followed the Helvetian tuff eruptions the area under consideration was an emergent land already.

At the Helvetian—Tortonian boundary appears the so-called pumiceous Middle Dacitic Tuff. Its upper member was redeposited in Early Tortonian time, as evidenced by its being mixed with Lower Tortonian volcanic detritus in a number of places. Its thickness is 70 m or so in the Fehérkő mine of the Csevice Valley and 60 m in borehole Tar-29. In the survey borehole Hasznos-l it is merely 39 m thick, but it should be taken into consideration that drilling was started from within this formation.

Controlled by rejuvenated fracture lines, Early Tortonian volcanism also yielded a considerable amount of lava. It brought about fissure volcanoes (e.g. Stremina crest), parasitic craters (e.g. the valley of Csörgő brook), minor volcanic cones (e.g. Kőerdő Hill) and, in some places, thin lava flows which were dismembered by subsequent erosion (e.g. on the southern side of the Farkaslyuk).

The volcanic cones consisting of andesites of dacitic nature (Óvár and Ágasvár), the subvolcanic, fresh, dark grey pyroxenic andesites and amafitic andesite masses and dykes appear to be of nearly equal age. In many cases a connection can be shown to exist between the subvolcanic products and the dykes. The dykes have not pierced the Upper Tortonian tuffitic limestones of Leithakalk facies anywhere; consequently, they are older than the Leithakalk. On the other hand, they are younger than the Lower Tortonian volcanic complex, because this is intersected by dykes.

The depressions of the resultant volcanic landscape were inundated by Late Tortonian sea, which thus produced the diatomite deposit of Hasznos and the tuffitic limestones of the Szalajka Valley near Tar, respectively. In addition, a faint eruption of pyroclastics should also be reckoned with, as evidenced by the rhyolite tuff bands intercalated within the sediments here. The thickness of the diatomite formation can be estimated at about 120 m in borehole Hasznos-4, that of the tuffitic limestones at about 70 m in borehole Tar-29. To the east of this area, the latter formation is only represented by thin rags which could escape erosion.

According to determinations by M. Bohn—Havas [Mezősi, 1963], the tuffitic limestones contain the following fossils: foraminifers — *Venus (Clausinella) basteroti* Desh., *Cyprina grinodica* Ben., *Pecten aduncus* Eichw. (fragment), *Phacoides (Linga) columbella* Lam., *Pitaria (Paradione) chione* Lam., *Arca sp., Lucina sp., Mactra sp., Meretrix sp., Tapes sp.,* Out of Bryozoans, *Vincularia sp.,* was recognized by Kolosváry.

According to J. Noszky Sr. [1927], at the confluence of Madarász and Csertő brooks there is a small patch of tuffaceous sediments of Leithakalk facies. They contain, beside *Amphistegina vulgaris* and *Heterostegina costata*, ill-preserved specimens of *Conus fuscocingulatus* Bronn., *Panopaea menardi, Cardium turonicum* May., *Pecten sp., Lucina sp., Serpula sp.* and *Dentalium sp.* At this locality, however, the tuffitic limestones are already redeposited, the autochtonous deposit being farther east.

After the withdrawal of Late Tortonian sea, this region also witnessed an intensification of erosion. On the norther side of the Mátra Mountains crest, talus fans were accumulated which presently vary between 10 and 30 m in thickness, locally attaining 95 m.

TECTONICS

The northwestern part of the Mátra Mountains is characterized by a faulted structure with chess-board-patterned fault-grabens and minor horsts. The present-day tectonic pattern of the investigation area is the result of repeated tectonic movements.

The earliest detectable tectonic movement of the area corresponds to the time of Late Helvetian volcanism. In fact, in the Ágasvár range, at the Szamárkövek and in Csörgő brook the Middle Dacitic Tuff is flanked by fracture lines of WNW—SES trend. Farther east, the Lower Tortonian volcanic complex lies, as exposed today, immediately on the Helvetian schliers, as the Helvetian volcanic complex and the Middle Dacitic Tuff are absent. Both these formations reappear farther east in the Polinka Valley along a fracture line of similar trend. The fracturing of the Helvetian lignite formation and of the schliers as well as their progressive plunging under the Mátra Mountains mass seems to correspond to the date of this faulting.

The second phase of differential movement coincided with the time of Early Tortonian volcanism when fissure volcanoes and minor parasitic craters were formed. Such a fissure volcano seems to be represented by the narrow crest running from Ágasvár towards Mátraszentiván, a volcanic range controlled by a fracture line of ENE—WSW trend. The similarly trending stretch of the valley of Csörgő brook and the WNW—ESE-trending valley of Narád brook also appear to belong to this category.

Lower Tortonian tectonic trends are also indicated by the dykes which partly coincide with the afore-mentioned directions and partly are of E—W or N—S trend. Both types of dykes represent lava masses of various sizes which have intruded into open fissures. The eventual vertical dislocations along dykes must be either pre- or post-volcanic, because dilatation joints, as a rule, cannot be supposed to be connected with, any major vertical displacement. The slight contact effects observable along the dykes were examined by Bognár and Póka [1964].

Post-Miocene crustal movements, whose manifestations can be distinctly demonstrated, were considerable, too. Whereas the earlier faults are oriented roughly ENE—WSW and WNW—ESE, respectively, the Miocene faults strike either NE—SW or NW—SE.

Most of the 150 deep boreholes drilled in the investigation area have reached the Burdigalian Lower Dacitic Tuff. With reference to this level, the size of displacement can be determined. On the basis of the plotted profiles, tectonic movements of various ages can be readily shown to have occurred. The gradual disintegration of the Helvetian schlier began as early as Late Helvetian volcanism.

Profile A—B of *Fig. 3* (70°—250°) extends from the borehole Nagybátony-109 drilled into the western bank of Kecskés brook, towards Ágasvár. Near Felső-Katalinbánya there is a horst-like hill. The eastern continuation of the andesite dyke exposed on the Csutaj can be encountered partly in underground workings, partly in minor dyke portions exposed to the surface. The volcanic rock hit at about 530 m in borehole Nagybátony-224 may be a portion of an andesite apophysis or of a subvolcanic body. Towards Ágasvár, the Helvetian schlier grows gradually thicker, to attain

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Fig. 2. Geologic structure of the area.

Legend: 1. Slope-detritus, alluvium; 2. Upper Pannonian sand; 3. Upper Tortonian tuffaceous limestone, diatomaceous earth; 4. Lower Tortonian andesite and andesite-tuff; 5. Dacite-tuff; 6. Helvetian andesite and andesite-tuff; 8. Helvetian coal series; 9. Burdigalian dacite-tuff; 10. Upper Oligocene; 11. Fracture-line; 12. Direction of geologic section.

→

Tar
Csevice brook
Szakad
Kőerdő peak
Csertő brook
Szalajka brook
A
Kecskés br.
Terendel brook
Szoros brook
Dobróka brook
Nyíres valley
D
Dobróka
E
F
Nagy-Bec meadow
Gombás peak
Mt. Várhegy
Csörgő brook
Mt. Hegyes
B
C
Polana valley
Csörgő brook
Hutahelyi brook
Nadá brook
LEGEND
1
2
3
4
5
6
7
8
9
10
11
12
A
B

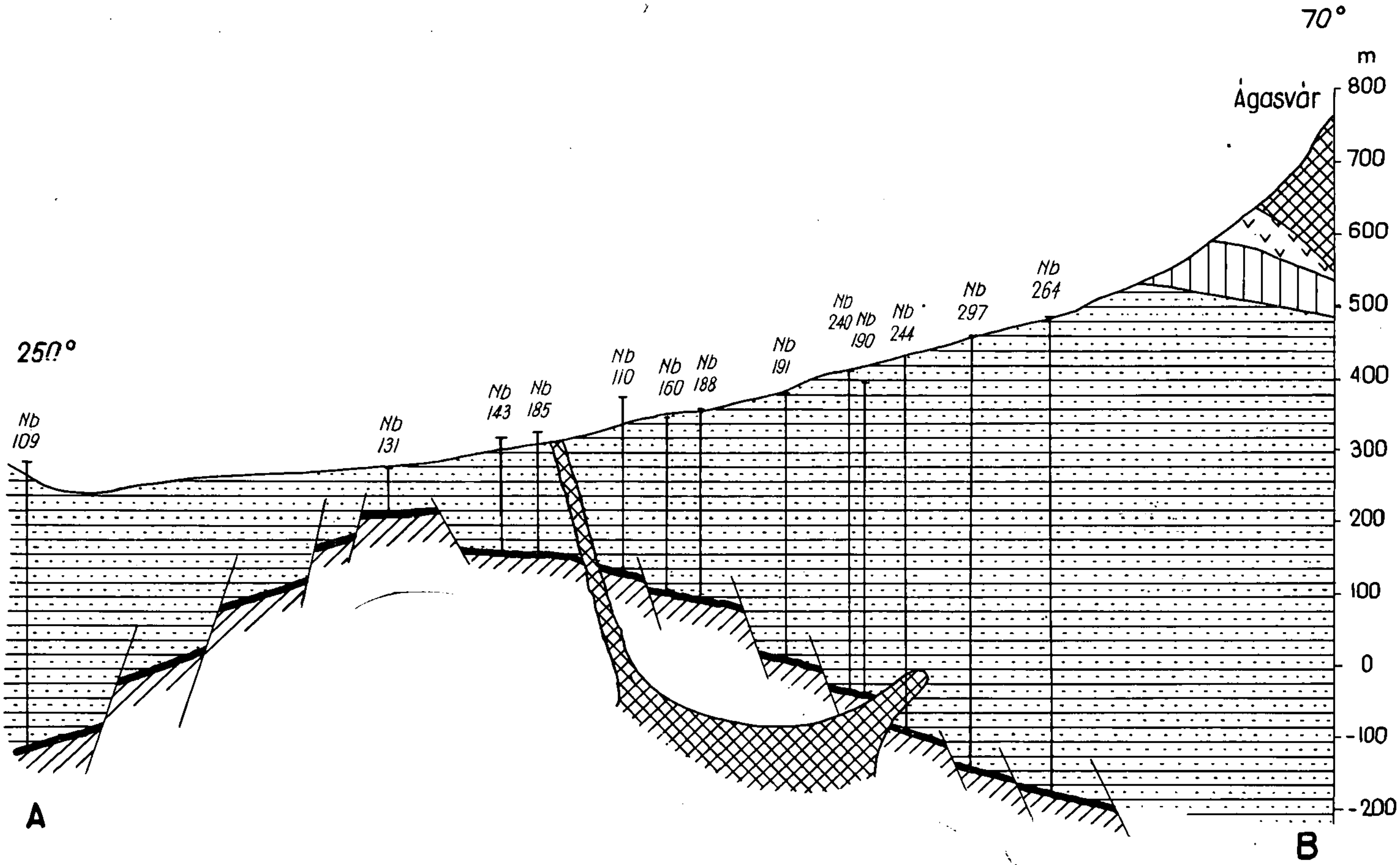

Fig. 3. Geologic section in the direction A—B on Fig. 2.

about 700 m in thickness beneath the Upper Helvetian agglomeratic andesite tuffs. The chess-board-patterned fracturing of the schlier and its southward tilting seem to be due to Latest Helvetian crustal movement. Later movements changed but little the position of the schlier. The faults strike at about 60° to 240° and 130° to 310°, respectively. Stratification planes dip usually southward at 6° to 12°.

Examination of geological structure along a profile (C—D, *Fig. 4*) of NE—SW orientation will also show the southward growth in thickness of the Helvetian schlier. In Tarkő brook — at about 330 m elevation a.s.l. — the Burdigalian Lower Dacitic Tuff is exposed. The borehole Nagybátony—105, drilled into the ridge between Sziget and Bükkös brooks, reached the Lower Dacitic Tuff (absolute elevation: +33 m) as high as at 413.4 m depth. Farther southwest, the tuff lies at nearly 550 m depth (absolute elevation: —80 m) in borehole Nagybátony—186.

The andesite dyke, exposed on the ridge running between Nagy Bec meadow and Tarkő brook in the eastern part of a profile (E—F) of approximately E—W trend *(Fig. 5)*, may be the off-shoot of a large subvolcanic body. This seems to be evidenced, on the one hand, by borehole Nagybátony—268 which, after crossing the lignite formation, was stopped at 443.6 m depth; on the other hand, by borehole Nagybátony—259 drilled into the bed of Sziget brook, which intersected the andesites twice, beneath 95 m of scree. Not far from here is the andesite dyke of the ridge between Sziget and Bükkös brooks. The andesite dyke, exposed in the bed of Bükkös brook, also seems to be connected with the afore-mentioned subvolcanic body. All of these andesite dykes strike NNW—SSE. The minor peaks of the Lower Dacitic Tuff along the profile are, for the most part, members of a southeastward horst range.

In Late Pliocene time a large NE—SW-trending fault graben was formed in the western part of the region. Whereas the Late Helvetian and Early Tortonian movements stroke at 60° to 240° and 130° to 310°, respectively, the strike of this fault graben corresponds to 30°—210°. The faults detected by Noszky Sr. [1927], Schréter [1940], and Szentirmai [1965], faults extending from Kőerdő Hill southwestwards, are only part of this system, as evidenced by the three boreholes, Tar-3, Tar-29, and Hasznos-4, drilled into the graben axis. Of these, the Lower Dacitic Tuff — exposed about one kilometer and a half farther east — was reached, at nearly 596 m depth, by only the borehole Tar-3 in the northeastern part of the fault graben. Drilled in the middle stretch of the graben, borehole Tar-29 cut tuffitic limestones under clayey talus down to 77 m depth. Underneath, 60 m of Lower Tortonian agglomeratic andesite tuffs followed. These were in turn underlain by the Middle Dacitic Tuff, again of 60 m thickness. The Helvetian schlier began at 447.5 m but was not cut through, as drilling was stopped at 675 m depth. However, considering the thickness of this formation in the near-by boreholes, the Burdigalian Lower Dacitic Tuff might be expected to occur here at about 820 m depth. The Upper Tortonian diatomite-bearing sequence, exposed near the Várhegy of Hasznos, was found to occur between 104 and 221 m in borehole Hasznos—4 (at the southwestern tip of the graben). This observation can be used for conclusions as to the height of faulting here. Since in the valley of Kövecses brook, near borehole Hasznos—4, the Middle Dacitic Tuff, marking the Helvetian—Tortonian boundary, is exposed to the surface, the fault plains must be steep. (Would this not be the case, so the Middle Dacitic Tuff would border on the older formation occurring on the eastern side of the fault.) Nota bene, borehole Hasznos—4 was stopped within Lower Tortonian andesite tuffs at 304 m depth.

One of the benches of this comparatively deep graben (about 450 m deep, as shown by drilling) is the fault detected by Noszky Sr. and Schréter. Parallel with it, there runs another, larger fault, indicated by Szentirmai, along which the Middle

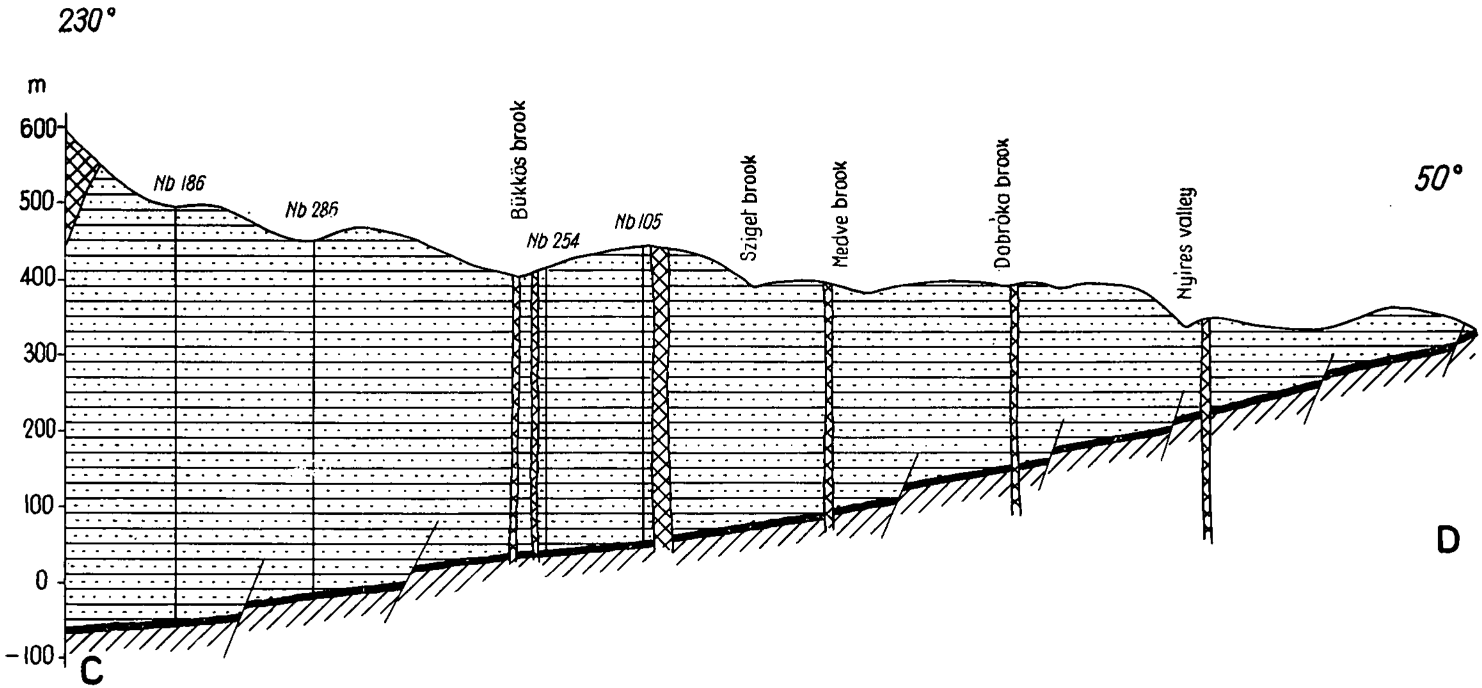

Fig. 4. Geologic section in the direction C—D, on Fig. 2.

Dacitic Tuff appears in the neighbourhood of the Helvetian schlier. It is in this large depression that the so-called „Szakadás gödre" was formed. Here only the Upper Helvetian agglomeratic andesite tuffs were intersected by borehole Tar—4.

In the vicinity of Gombás Hill this graben was filled up in Late Pliocene to Early Pleistocene time. The talus deriving from Mátrakeresztes is constituted partly by clay-mineralized amafitic andesites of onion-shaped (curbicortical) weathering, partly by dark grey, less weathered pyroxenic andesites. In addition to them, debris of jasper, opal, chalcedony, and veined quartzite are common, the last of which sometimes carry parasitic baryte plates. These are likely to represent residues of erosion of the baryte veins occurring in the vicinity of Mátrakeresztes.

Similarly in Late Pliocene time, a fault system of approximately NW—SE strike was formed. As the most eloquent example of it, the environs of Hegyes Hill may be quoted. The geological survey borehole on the southern slope of Hegyes Hill cut first the Middle Dacitic Tuff and then penetrated into Helvetian schlier at 173 m depth. However, at about 100 m south of the bore-head, the schlier is already exposed to the surface. This part of the Kövecses Valley has been controlled by this fault. The valley stretch by the Várhegy is also connected with the same fault. Between boreholes Hasznos—2 and Hasznos—3 there is a difference of 96 m in the hypsometric position of the Burdigalian Lower Dacitic Tuff, a phenomenon which is also due to a fault of NW—SE trend. The outcrop of the Middle Dacitic Tuff on the eastern slope of Gombás Hill is also fault-controlled, since Helvetian schlier lies close to it on the eastern side.

The NE—SW-trending fracture lines in the vicinity of Fehérkő mine near Tar village are known from earlier literature data [Noszky Sr., 1927., Schréter, 1940., Kubovics, 1963].

CONCLUSIONS

The investigation area lies in the northwestern Mátra Mountains. Uncovered by drilling in the northwestern foreland of the Mátra Mountains, the oldest formation of the area is the sedimentary sequence of the Rupelian stage. This is overlain first by Upper Oligocene, then Burdigalian sequences. The footwall of the Helvetian lignite formation is constituted by the Lower Dacitic Tuff. After lignite deposition the rate of subsidence was accelerated. The lignite-bearing sequence is overlain, with intermediary of Chlamys sands or without them, by marly, micaceous siltstones (schliers), whose thickness may locally approach to 700 m. This formation is commonly poor in fauna. In Helvetian time, regression was coupled with andesitic volcanism. The resultant pyroclastics, appearing at the Helvetian—Tortonian boundary, have been termed the Middle Dacitic Tuff. Early Tortonian volcanism in this area produced fissure volcanoes, parasitic craters, minor volcanic cones, and thin lava flows. All of the subvolcanic bodies, inclusive of dykes, are of subequal age. Certain depressions of the resultant volcanic landscape were invaded by Late Totronian transgression which brought about tuffitic limestones in the Szalajka Valley near Tar village and diatomite accumulations by Hasznos. After regression of Late Tortonian sea, erosion processes revived here too.

The earliest detectable tectonic deformation in the area under consideration corresponds to the date of Late Helvetian volcanism. It is characterized by WNW—ESE and ENE—WSW trends. The second tectonic phase coincided with Early Tortonian volcanism, when fissure volcanoes were formed. In addition, valley tracks have partly been controlled by the same movements. Trends aggree with those of the first

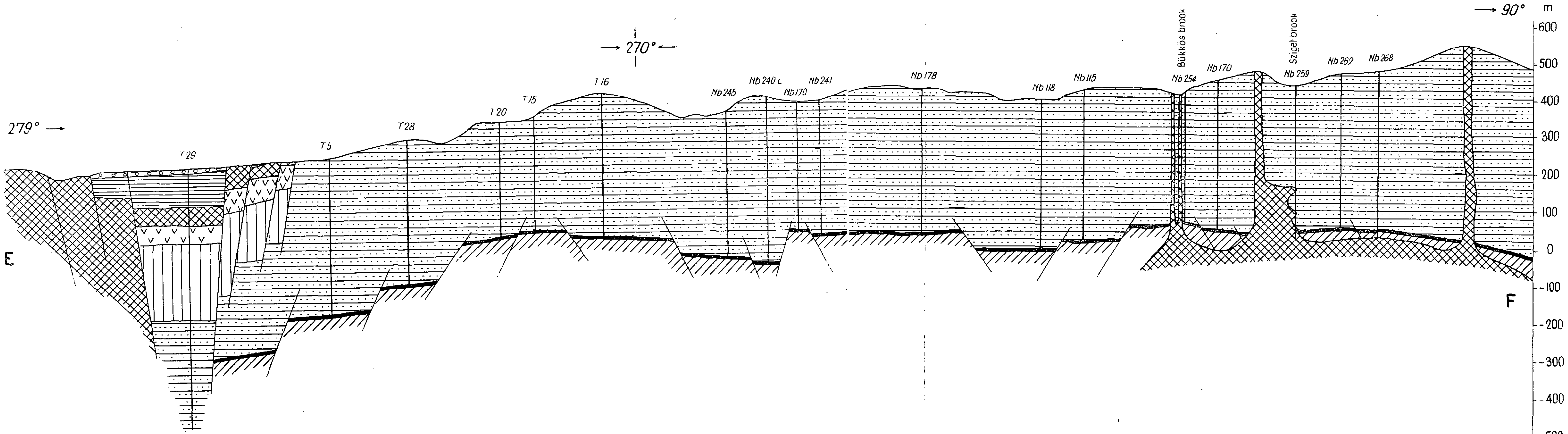

Fig. 5. Geologic section in the direction E—F on Fig. 2.

phase. Early Tortonian tectonic trends are also indicated by the strike directions of dykes intruded into dilatational fissures. Post-Miocene movements, of NE—SW and NW—SE trends, were also remarkable, as evidenced by the formation of a step-faulted, deep graben characterized by level differences of nearly 450 m and by steep fault planes.

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SEDIMENTOLOGICAL INVESTIGATIONS OF UPPER PANNONIAN AND PLEISTOCENE DEPOSITS IN THE NORTHEASTERN GREAT HUNGARIAN PLAIN

B. Molnár

INTRODUCTION

The northern and northeastern border of the Great Plain's Pliocene basin is made up of Miocene volcanic rocks in most places. In the north the Mátra Mountains andesites, the Bükkalja (foreland of the Bükk Mountains) rhyolite tuff mantle and the complex eruptive mass of the Tokaj Mountains form this border, in the east it is constituted by the andesite volcanic range punctuated by the Vihorlát, the Kőhát, the

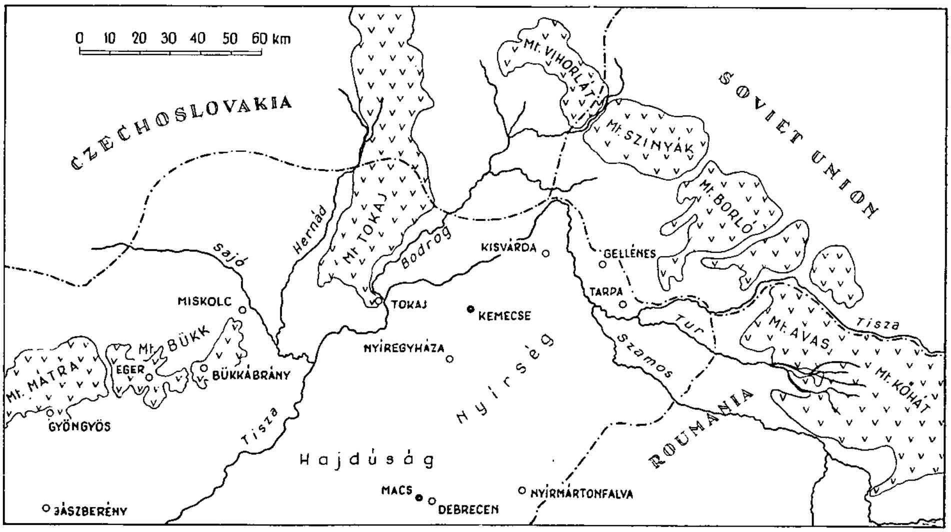

Fig. 1. Layout

Avas, and the Gutin Mountains *(Fig. 1)*. The volcanic formations continue towards the basin interior, as evidenced by gravimetric anomalies, by the andesite "islands" of Tarpa and Zemplén, and by the deep boreholes drilled in the Nyírség and Hajdúság areas.

The thickness of the volcanics decreases southwards. The Kisvárda borehole has penetrated into a 1050-m-thick rhyolite tuff complex after having traversed 1200 m of younger sedimentary deposits. The Nyíregyháza borehole, after crossing 1150 m of sediment, has cut 1425 m of Miocene volcanics without getting out of it. The Gelénes borehole has uncovered a volcanic complex of similar thickness [J. MOLNÁR, 1965]. Farther south, the borehole drilled at Nyírmártonfalva has found the same formations to be not thicker than 445 m [L. KŐRÖSSY, 1956, 1957, 1962].

In the northeastern part of the Great Hungarian Plain, inside the frontier, the Tortonian marine and Sarmatian brackwater sediments overlie volcanic products averaging 100 to 200 m in thickness. These may be intercalated by the afore-mentioned sediments or even be present in form of isolated patches only.

The total thickness of the Pliocene sediments, however, even exceeds 1000 m. Their stratigraphic subdivision has not been solved satisfactorily in all of the occurrences known. However, the general scheme adopted is to class their lower — thicker — part as Pannonian which is further subdivided into substages. The upper part is represented by the Upper Pliocene ("Levantine") sequence.

The Lower Pannonian oligohaline, lacustrine sediments are even known from outcrops in the Bükkalja and the Hernád valley. The Bükkalja Lower Pannonian sequence, transgressive on different members of the Miocene rhyolite tuffs, include in their basal layers the redeposited materials of their foot-wall, being represented by whitish, yellowish and greenish, tuffaceous sands and clays. However, the bulk of the sequence is constituted by yellow sands, sandstones, yellow and gray clays, calcareous clays, and clay-marls accompanied, in some places, by fine-grained gravels and granules as well as by thin lignite stringers. The Lower Pannonian outcrop of the Hernád valley consists of alternating sands and clays [Z. SCHRÉTER, 1939; K. BALOGH, 1964; K. BALOGH—A. RÓNAI, 1965].

In the Lower Pannonian, represented by basin facies, the predominant sediments are grey and bluish-grey clays and clay-marls with sand and sandstone lenses of different size, though commonly a few metres thick only. The thickness of the basin-facies Lower Pannonian attains 1000 m in the northern Great Hungarian Plain, along the Tisza river; farther east of this line, these sediments thin out to be reduced to 97 m at Nyíregyháza, to 68 m at Gelénes, and to 82 m at Nyírmártonfalva [E. R. SCHMIDT, 1939; L. KŐRÖSSY, 1957; L. DUBAI—K. JAMNICZKY, 1961; V. DANK, 1961].

The Mátra—Bükkalja sector of the Upper Pannonian freshwater (lacustrine, paludal) basin-filling sediments is constituted by greenish-grey and grey clays, clay-marls, grey to yellow sands as well as sandstones. Within these alternating sediments, the sandy members are predominant. The sequence is characterized by the intervention of numerous lignite seams which often attain several metres in thickness. The outcrop lying east of the Hernád valley shows similar lithological composition, the only difference being the absence of lignite.

On the basin border the Upper Pannonian formations transgress the Lower Pannonian and overlie directly the Miocene rhyolite tuffs.

The basin-facies Upper Pannonian of the northeastern Great Hungarian Plain is represented by a frequent alternation of sands and sandy clays locally interrupted by marls, lignite stringers, and calcareous concretions. The thickness varies between 500 and 1000 m.

Difficult to distinguish from the Upper Pannonian is the Upper Pliocene ("Levantine") fluviatile, paludal, continental sequence characterized by the scarcity or total absence of fossils, and lithologically by variegated (streaked, mottled), ill-stratified or fine-sandy clays and light grey, fine-grained sands with plenty of calcareous concretions [M. SZÉLES, 1965). According to some workers [M. ERDÉLYI,

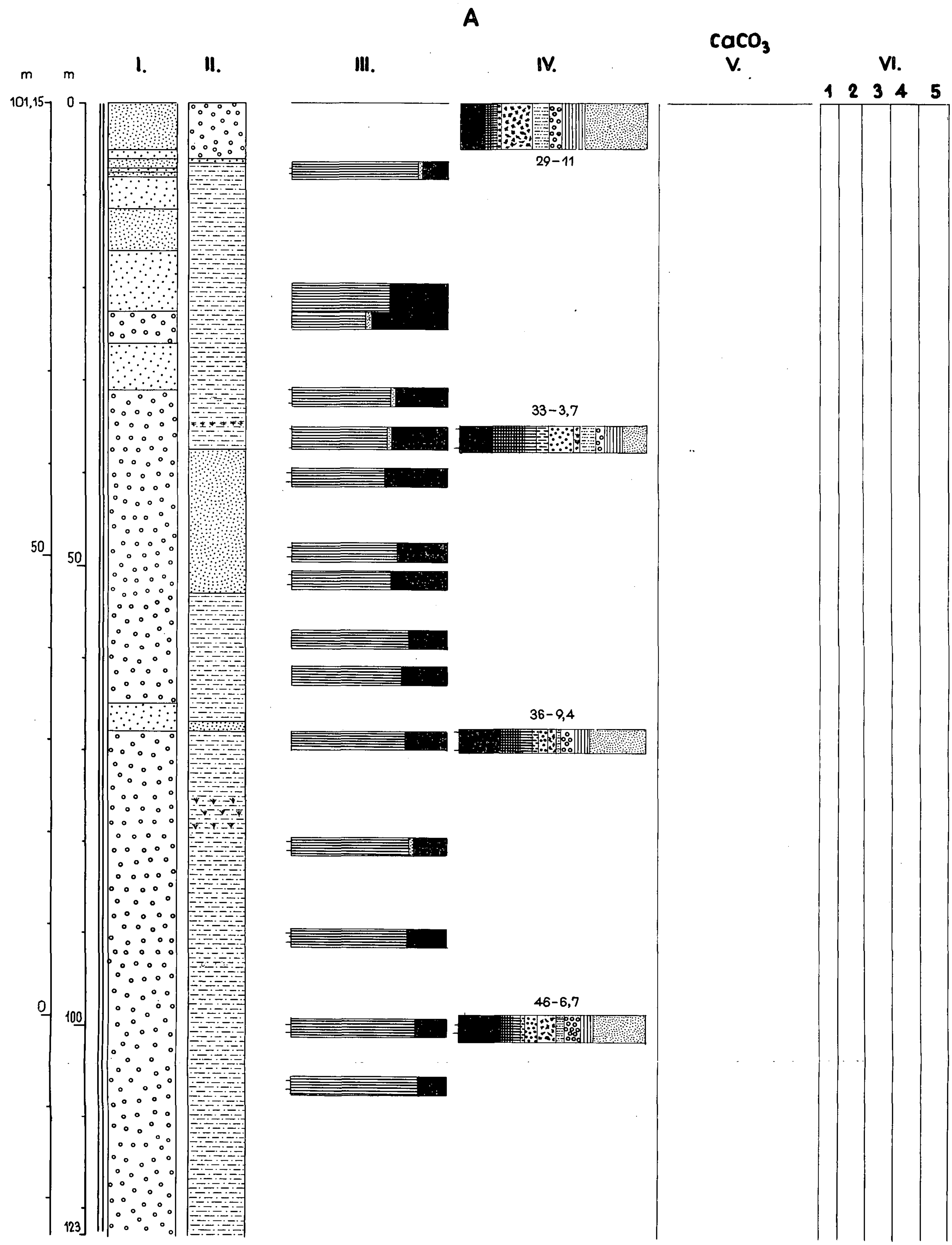

Fig. 2. Lithologic log of the Kemecse borehole, interval of 0 to 123 m.

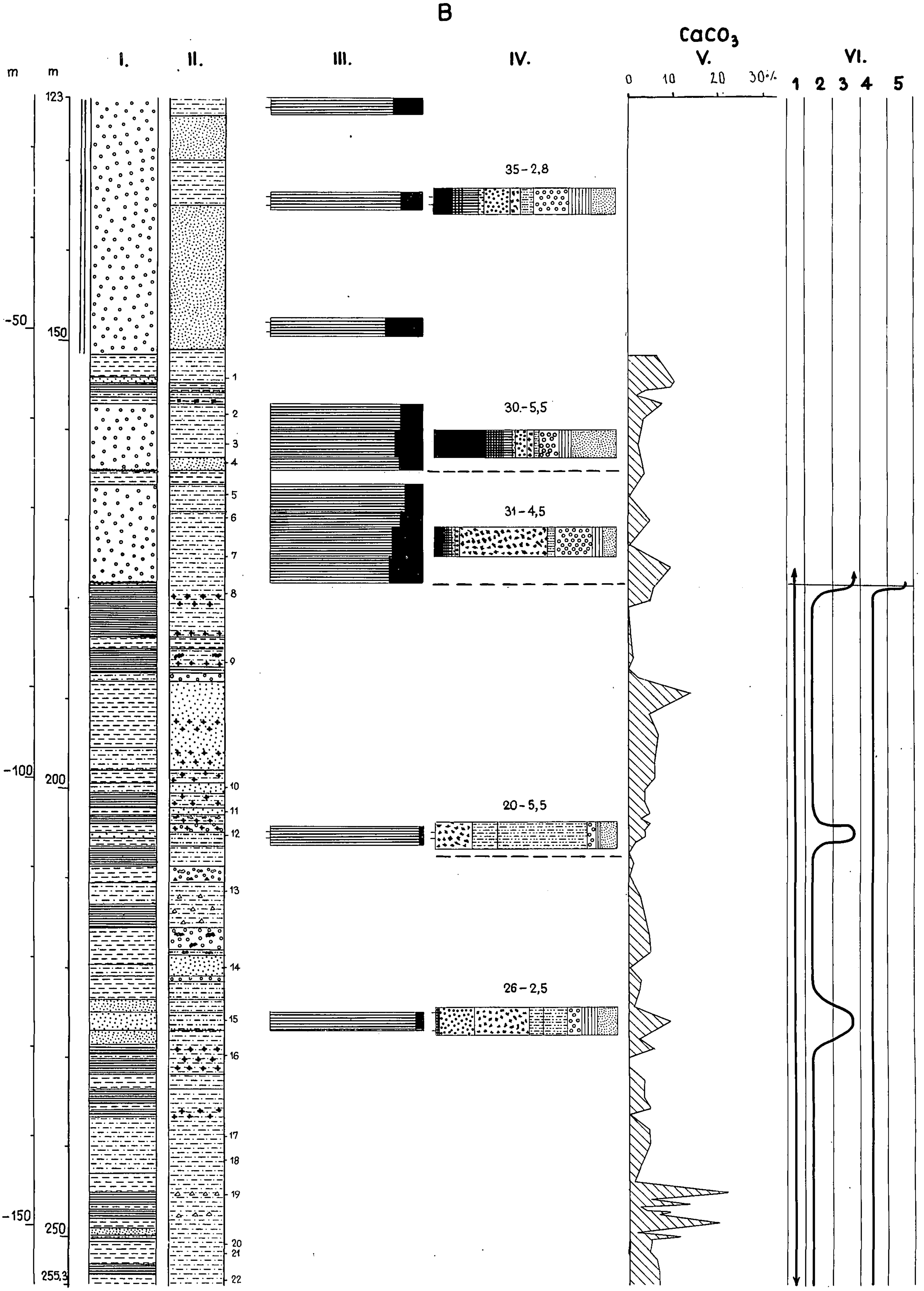

Fig. 3. Lithologic log of the Kemecse borehole, interval log of 123—253,5 m.

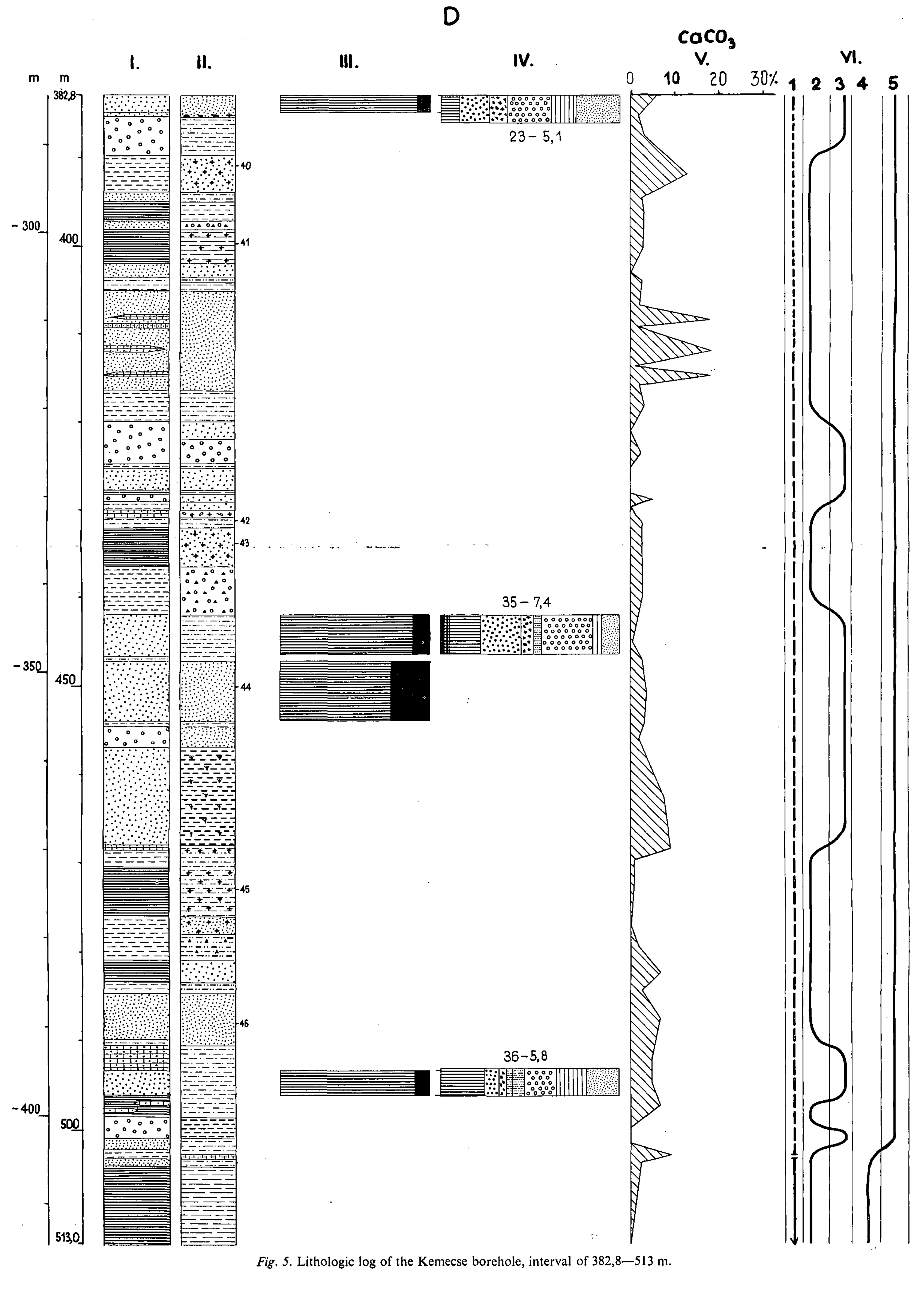

Fig. 5. Lithologic log of the Kemecse borehole, interval of 382,8—513 m.

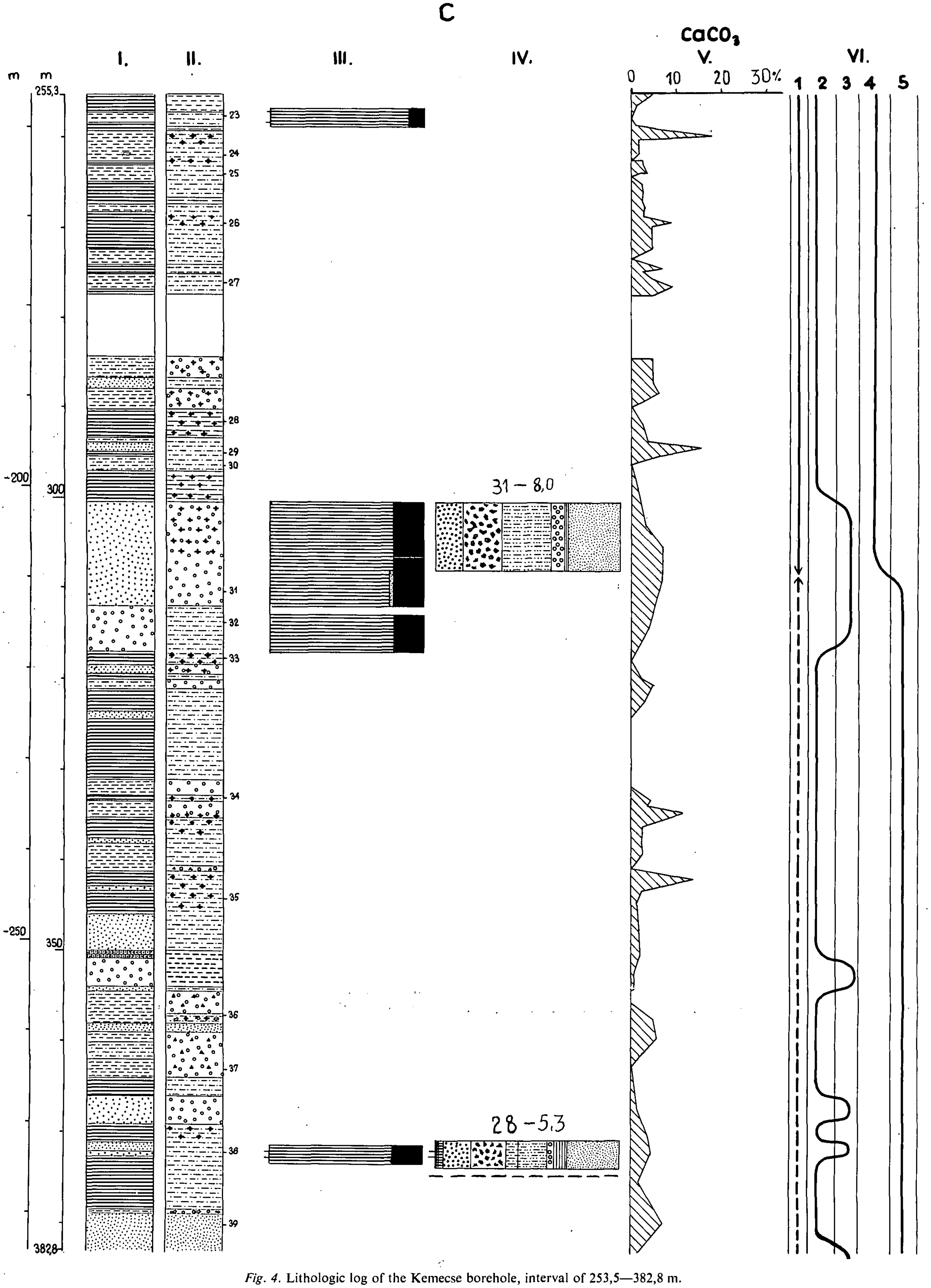

Fig. 4. Lithologic log of the Kemecse borehole, interval of 253,5—382,8 m.

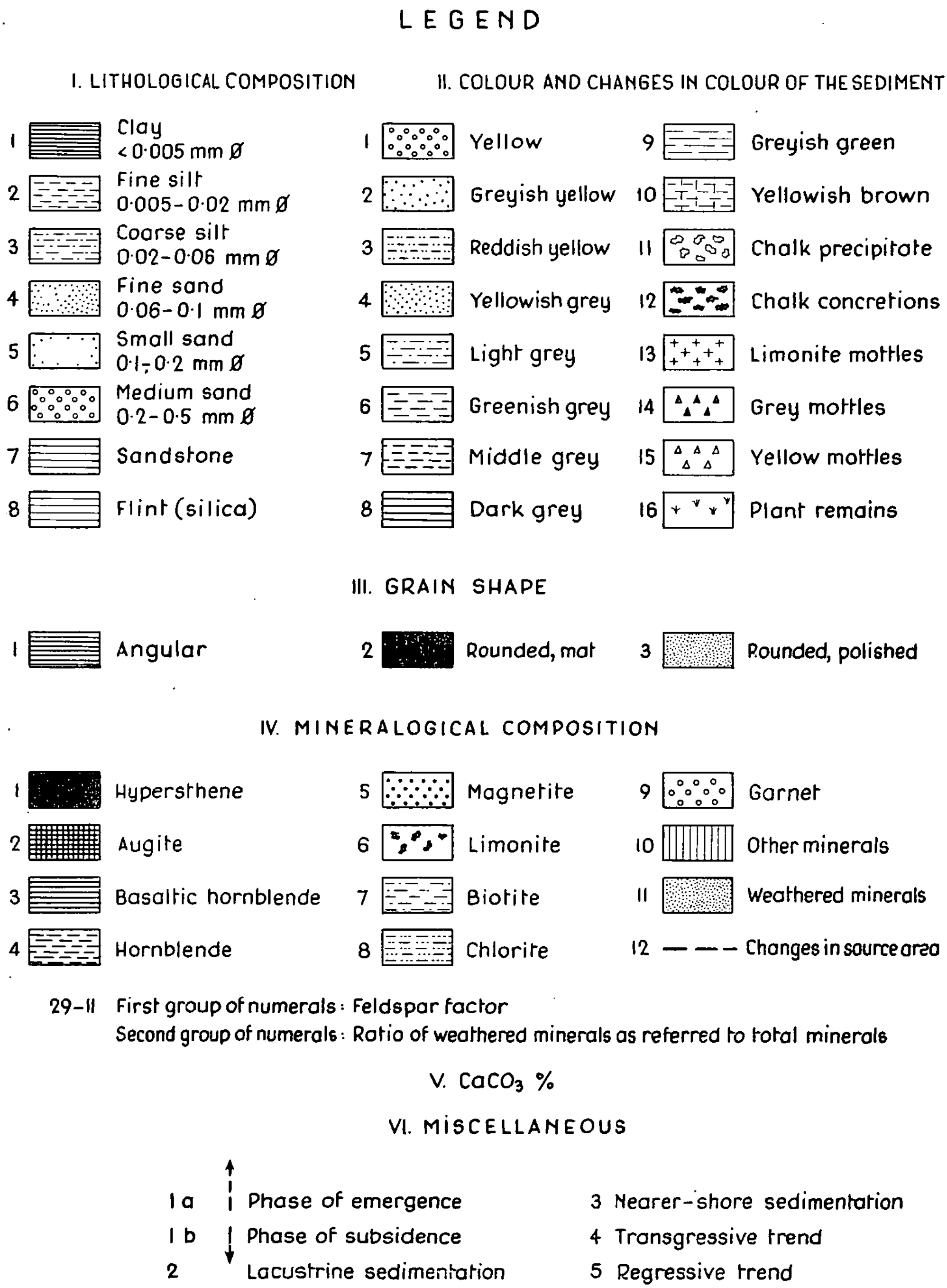

Fig. 6. Legend to *Figs. 2* to *5*

1960; L. KŐRÖSSY, 1962], this formation would be absent in the norther Great Hungarian Plain, while others believe it to be represented by a thickness of 200 to 400 m [I. DOBOS, 1965; M. SZÉLES, 1965].

In post-Pliocene time the northeastern Great Plain area was completely emerged and affected by considerable down-drops (sudden subsidences). The Pannonian ridges which remained emergent between the subsided parts are no longer reflected by present-day morphology, for the 50 to 100 and even 250 m level differences have been planated by the wind-blown and water-transported sediments of the Quaternary period. South of Nyíregyháza, the Pannonian surface shows particularly high and varied relief. The Hajdúság is formed by a Pannonian plateau of higher position with regard to its neighbourhood, where the Pannonian sequence can be reached as high as about 10 m below the surface [M. ERDÉLYI, 1962; B. MOLNÁR, 1966].

The 50- to 250-m-thick Quaternary sequence, represented predominantly by fluviatile sediments, shows a reduction of grain size from the mountain frame towards the centre of the Nyírség area [J. URBANCSEK, 1965]. In the Bodrog-köz area the fluviatile deposits are still rather gravelly, but as one proceeds southwards, the sand fraction will gain predominance, being intercalated by several silt layers.

According to J. SÜMEGHY [1944, 1955], the Nyírség Pleistocene would suggest a comparatively late subsidence. Consequently, the majority of the alluvial fan sequence would represent the second half of the Pleistocene.

The 20- to 25-m-thick Pleistocene sequence of the Nyírség is constituted by uppermost Pleistocene loesses and wind-blown sands and Holocene eolian sands [A. RÓNAI —L. MOLDVAI, 1966].

In 1958 and 1959 the Hungarian Geological Institute had a borehole drilled, by core-drilling for the most part, at Kemecse and Macs in the northeastern Great Hungarian Plain, a measure which aimed at a detailed investigation of the poorly known Upper Pannonian and Pleistocene basin facies and at the exploration of the possibilities for tapping artesian waters *(Fig. 1)*.

The hydrogeological evaluation of the Kemecse borehole was performed earlier, by M. ERDÉLYI [1960]. However, the careful lithological processing of the cores has been delayed to the time of the present study. The writer of the present paper should like to fill this gap, to enhance a better understanding of the facies of the Upper Pannonian basin sediments, to provide a contribution to the problem of the absence or presence of the Upper Pliocene in the northeastern Great Hungarian Plain, to promote the lithological locating of the Pliocene—Pleistocene boundary, and, finally, to determine the sources (origin) of the Upper Pannonian and Pliocene sediments.

Within the 513 m log of the Kemecse borehole, a substantial lithological change can be observed at 177 m. As shown by the interpolated data of L. KŐRÖSSY [1962], the total thickness of the Upper Pannonian sequence may be 600 to 800 m. The lower 336 of the Kemecse borehole has uncovered about the half, and surely one-third, of this. The uncovered interval is exactly that which comprises the major part of the sedimentary sequence of the final regression of the Upper Pannonian inland sea which used to cover the entire Great Hungarian Plain area.

A considerable part of the lithological log and of the analyses have been illustrated in *Fig. 2* to *Fig. 5* (in four divisions: A, B, C, D). In each of these, the absolute altitude and the depth of drilling have been indicated on the left side of the lithologic log. The rest of the signs have been given in the legend *(Fig. 6)*. In the figures the finer and darker sediment has been indicated by a denser hachure, the coarser and lighter one by a wider-spaced hachure. The numerals beside the IInd column of the figures are identical with the serial numbers of Table 1.

Grain composition and its statistical evaluation

According to grain composition, the borehole can be divided into two substantially different parts: *a)* in the 177 to 513 m interval the sediment is represented predominantly by clays and silts with somewhat less fine sand, i.e. sediment of finer grain fraction, *b)* the 0 to 177 m interval is constituted by medium sands. The medium sands of the 157 to 177 m interval include some 10 to 15% of coarse sand, too: this is the coarsest part of the entire lithological log *(Fig. 2* to *Fig. 8)* (Samples *1* to *46*).

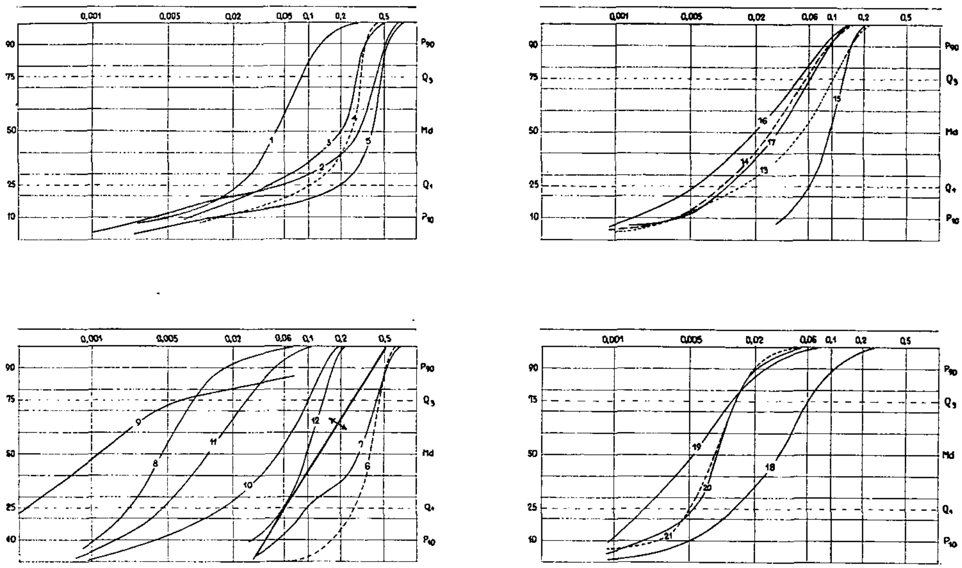

Fig. 7. Granulometric curves of the samples from the interval 0 to 252,3 m

The lithological statistical values of the examined strata have been presented in *Fig. 9* and Table 1. Although not all of the strata have been analysed in detail, the analyses have been extended to all lithological types. The analysed samples are rather evenly distributed within the profile, so that the variation of the results are suitable for the characterization of the sedimentation processes. The 0 to 150 m interval of the borehole was drilled with full hole, but bottom hole flushing was applied at 50 cm intervals. Consequently, the taken samples cannot be analysed granulometrically.

The value of sorting *(So)* varies within a very wide range (1.15 to 5.25). However, stretches with well-sorted and with more poorly sorted sediments can be selected. For instance, a good sorting has been found in the 372—513 m (Samples *38—46*) and the 0—255.3 m (Samples *1—22*) intervals, while the sediments of the 253.3—372 m (Samples *23—37*) interval have proved to be less and rather variably sorted. The last-mentioned phenomenon reflects the unsteadiness of sedimentation. In this interval the other sedimentological characteristics also change.

The value of curtosis *(K)*, indication the ratio of 50% to 90% of the curve, varies between 0,18 and 0,38. In the sequence below 245 m it has a lower value, above this level, a higher value. This means that in the lower part of the sequence the fluctuation

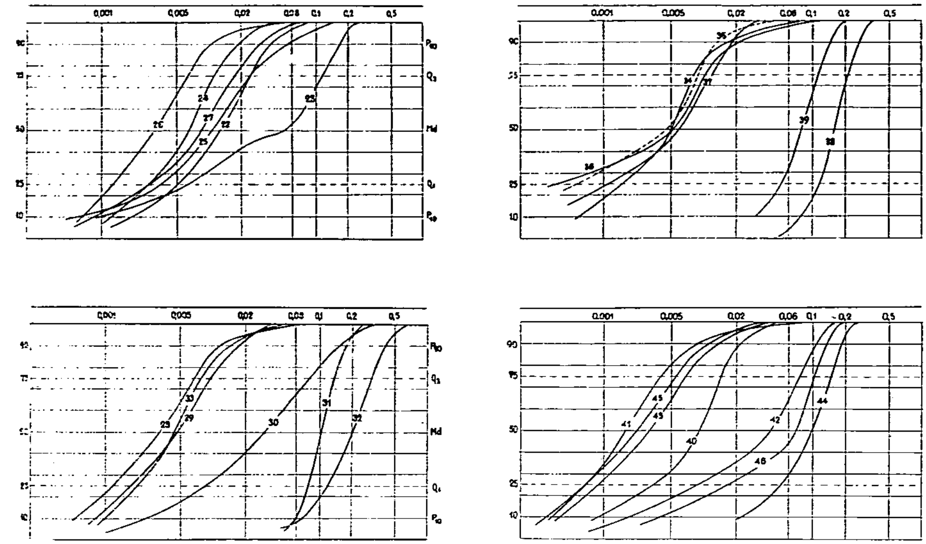

Fig. 8. Granulometric curves of the samples from the interval 254—490 m

of deposition energy was smaller, in the upper part (the complete regression sequence of the Pannonian inland sea and the Pleistocene fluviatile sequence) it was greater.

The skewness *(Sk)*, indicating the asymmetry of grain size with regard to the average value and showing the deviations of the changes of deposition energy from the average energy, varies parallel to the *K* value. Accordingly, below the 245 m level

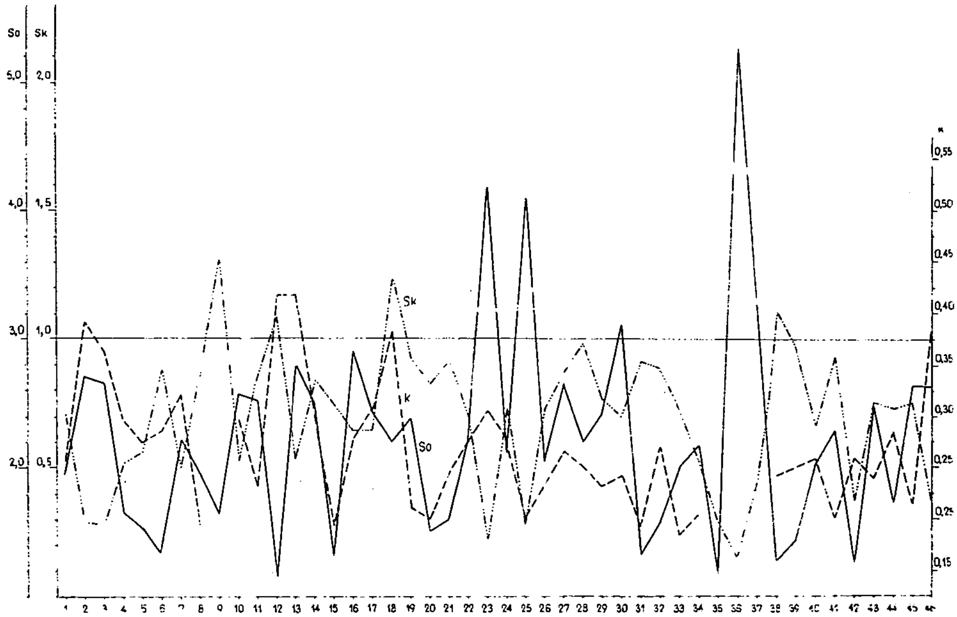

Fig. 9. Lithological statistics of the analysed samples (the numerals equal the serial numbers on the side Column II., *Figs. 2—5* as well as those of Table 1).

it is usually smaller, while above it greater values can be observed (within the 0.23 to 1,33 range, just like the amplitude of K variation (*Fig. 9*, Table 1).

Consequently, on the basis of grain composition and its statistical characteristics, three intervals of different sedimentation conditions can be distinguished in the Kemecse borehole log: *a)* the interval below 245 m characterized by a relatively steady, balanced sedimentation, *b)* the 177 to 245 m interval characterized by more unsteady, more rapidly changing conditions, and *c)* the 0 to 177 m interval suggesting sedimentation conditions substantially different from the former two. Intervals *a* and *b* are the result of continuous sedimentation, while interval *c* overlies unconformably the *a* and *b* sequences.

Grain shape

The shape of sand grains was examined by the method of A. CAILLEUX [1952, 1961]. Thus three grain types could be distinguished: *a)* gritty-splittery, unrounded grains, *b)* rounded, bright-faced grains transported by water for more than 400 km distance, *c)* rounded, dull grains characteristic of wind-blown sands. The method applied is a statistical one. Accordingly, the problem of origin is determined by the predominant grain type within the sediment.

In the Upper Pannonian sequence of the Kemecse borehole, gritty-splittery grains of *a* type were found to predominate; in addition to them, a few slightly rounded, dull grains have also been found. Although confined to 308 m depth, 1 to 2% rounded, bright grains have also been encountered.

In all of the Pleistocene samples the water-transported grains are predominant, though in the 15 to 52 m interval the ratio of the rounded, dull, wind-blown grains is comparatively higher. Incidentally, some water-transported, bright, rounded grains can also be encountered. These must have derived from the redeposition of older sediments, for the length of the rivers of the northeastern Great Hungarian Plain does not attain 400 km. Hence, these grains may derive from the coasts of the Pannonian inland sea (*Fig. 2* to *Fig. 5*, IInd column, and *Fig. 6*, IIIrd column).

Mineralogical composition

1. Heavy minerals. In the Upper Pannonian sequence of 336 m vertical range the heavy mineral composition changes thrice, in the 177 m of Pleistocene sediment, once. Hence, from bottom to top, the following types can be distinguished:

a) The sand samples of the interval, ranging from 372 m down to the bottom of the borehole, are dominated by basaltic hornblende and garnet, while magnetite and limonite show an upward increase. Two samples from this interval have a considerable tourmaline content (*Fig. 5*, Table 2, Samples *12* to *14*).

b) The samples of the 225—372 m interval can be distinguished from the underlying sediment by their higher chlorite and limonite contents, while basaltic hornblende and tourmaline vanish or are reduced to insignificant amounts in them. The frequency of garnet also diminishes. Magnetite keeps on being an important component (*Fig. 3* and *Fig. 4*, Table 2, Samples *9—11*).

c) The Upper Pannonian sample from 204 m is conspicuous for its striking chlorite (50%) and biotite (13,4%) contents. However, magnetite disappears and it is alone limonite that attains a 20% ratio even here (*Fig. 2*, Table 2, Sample *8*).

d) The Pleistocene sample from the 171—174 m interval shows a mineralogical composition which is intermediary between the underlying Upper Pannonian and the overlying younger Pleistocene sediments. Limonite is abundant, hypersthene augite and garnet appear (*Fig. 2*, Table 2, Sample 7.).

e) Finally, the sediments of the 0 to 163 m interval are characterized by a composition corresponding to that of the recent sediments of the Tisza river: hypersthene, augite, basaltic hornblende, magnetite, garnet, and limonite [B. MOLNÁR, 1964].

2. Light minerals. First of all, the quantitative changes of feldspar grains have been studied on the basis of the method developed by A. CAILLEUX [1965]. Accordingly, different causes may provoke the enrichment of feldspar (cold climate, proximity of eruptive rocks). However, the decrease of the amount of feldspar is always due to the intensification of chemical weathering (and *not* to physical disintegration). According to CAILLEUX, a sediment characterized by a feldspar index of 10 to 80 is considered feldspar-rich, while one having an index of 4—5 is considered feldspar-poor.

Although the material of the Kemecse borehole can be declared feldspar-rich on the whole, yet various intervals of different feldspar content can be distinguished within the stratigraphic column of the borehole. In those two samples taken from the basalt stretch of the borehole in which the ratio of basaltic hornblende attained 19 to 23%, the feldspar index too is 35 to 36 *(Fig. 5)*. The feldspar index decreases with the upward decrease of the ratio of magmatic minerals, being as low as 25 to 30 in the rest of the Upper Pannonian sequence. In the lower sample of the Pleistocene it is 31, but it increases upwards with increasing feldspar ratio. Its value of 29 to 46 seems to vary parallel with Pleistocene climatic changes (*Fig. 2—Fig. 5*, IVth column).

The ratio of the weathered and coated grains to the total of the mineral grains (2,5 to 11%) has also been examined, but this does not show any striking regularity.

Carbonate content

As compared to the rest of the Great Hungarian plain, the carbonate content is low throughout the log, being 4 to 7% on the average. The highest value does not attain 25% either.

THE HISTORY OF ACCUMULATION AS EVIDENCED BY THE KEMECSE BORE LOG

1. *Upper Pannonian sequence:* in the 253.5 to 372 m interval most of the lithological characteristics are other than below and above this interval. Consequently, a substantial lithological change has taken place within this interval. That this must have been due to tectonic movement is evidenced by the change of source area. Thus the 338-m-thick Upper Pannonian section can be subdivided into two sedimentation phases at 372 m.

a) Within the 253.5—372 m interval it is difficult to draw any exact limit of overall change in sedimentation, for this change was gradual.

Still, the lower 202 m stretch (between 300 and 502 m) can be regarded as a minor regression phase on the basis of its position within the entire geological section (*Fig. 3* and *4*; VIth column). At 502 m the borehole has ended in clay. From this level up to 300 m the fine sands also play an important role. Between 372 and 382 m, as shown above, the source area has also changed. Underneath, the presence of basaltic hornblende indicates the redeposition and reworking of the late volcanic materials of the marginal zone. The presence of garnet may suggest the redeposition of garnet-bearing older sediments. In the stretch above 372 m the disappearance of basaltic hornblende and the considerable amount of chlorite prove the intensification of the erosion of sediments. This is also confirmed by the feldspar index which has the highest

Table 2

Heavy mineral composition of the Upper Pannonian and Pleistocene sands of the Kemecse borehole

Number	Depth m	DOMINANTLY MAGMATIC MINERALS									DOMINANTLY METAMORPHIC MINERALS										OTHER MINERALS				Total quantity of minerals in the examined fraction	Dominant grain diameter mm	Age
		Hypersthene	Other rhombic pyroxenes	Augite	Diopside	Basaltic-hornblende	Magnetite	Biotite	Apatite	Zirkon	Chlorite	Tourmaline	Epidote	Zoizite	Rutile	Hornblende	Actinolite-tremolite	Garnet	Staurolite	Cyanite	Calcite-dolomite	Limonite	Other micas	Weathered minerals			
1.	0—5	12,3	—	4,5	0,7	1,3	1,3	0,7	—	—	9,7	1,3	—	—	—	—	—	6,5	—	—	—	18,2	9,8	33,7	0,9	0,06—0,1	Pleistocene
2.	35—37	17,4	1,5	16,9	0,5	6,6	13,4	0,5	3,1	—	8,7	1,5	—	—	—	6,7	—	4,6	—	—	—	3,6	2,6	12,4	0,9	0,2—0,5	
3.	68—70	22,2	2,8	10,4	0,7	7,0	4,2	—	—	—	2,1	0,7	1,4	—	—	3,5	0,7	8,3	0,7	—	—	4,8	—	29,5	3,6	0,2—0,5	
4.	100—101	22,4	0,7	6,3	1,4	4,3	6,8	—	2,1	—	4,9	—	—	—	0,7	2,1	0,7	9,1	0,7	—	—	9,8	0,7	27,3	2,1	0,2—0,5	
5.	134—135	10,2	1,1	5,4	3,2	8,7	14,6	—	3,2	—	7,2	1,1	—	—	1,6	3,2	1,1	19,4	—	0,6	—	5,9	1,1	12,4	3,5	0,2—0,5	
6.	160—163	28,5	1,3	9,5	0,6	5,1	6,3	—	0,6	—	2,5	0,6	—	—	—	1,9	3,3	11,4	—	—	—	3,7	0,6	24,1	1,5	0,2—0,5	
7.	171—174	4,4	1,3	3,8	0,6	1,3	1,3	—	—	0,6	4,4	0,6	1,3	—	—	1,3	—	20,7	—	—	—	50,2	—	6,9	1,5	0,2—0,5	
8.	204,8—205,5	—	—	—	—	—	—	13,4	—	—	50,1	—	1,3	—	—	—	—	4,9	—	1,3	—	20,4	—	8,6	0,5	0,1—0,2	Upper Pannonian
9.	225,5—226,0	0,7	—	1,4	—	—	19,4	7,6	4,9	—	13,2	1,4	0,7	—	0,7	—	—	8,3	—	—	—	31,3	—	10,4	1,0	0,1—0,2	
10.	300,5—307	—	—	—	1,1	—	14,7	—	—	—	26,2	—	—	—	—	—	—	7,9	—	1,1	—	21,5	—	27,5	0,2	0,06—0,1	
11.	371—372	0,6	—	1,2	—	1,2	17,9	5,4	1,2	—	16,0	1,8	—	—	1,8	—	—	3,6	—	—	—	19,6	2,9	26,8	0,6	0,1—0,2	
12.	382,8—384,8	—	—	—	0,9	9,7	17,5	—	—	—	—	6,1	2,6	0,9	—	—	—	25,4	—	—	—	9,7	3,5	23,7	1,6	0,06—0,1	
13.	442—446,7	1,8	—	1,8	—	19,0	22,6	—	1,8	—	4,7	1,8	—	—	—	—	0,9	29,6	—	—	—	6,6	—	9,4	0,4	0,1—0,2	
14.	493,4—498,6	—	—	—	—	23,2	9,7	3,5	—	—	7,0	5,3	2,6	—	—	—	—	17,8	—	—	7,0	4,4	—	17,9	1,1	0,1—0,2	

TABLE 1

Lithologic statistics of the studied Upper Pannonian and Pleistocene material of the Kemecse borehole

Number	Depth m	P_{10}	Q_1	Md	M	Q_3	P_{90}	So $S_o = \sqrt{\frac{Q_3}{Q_1}}$	K $\left(K = \frac{Q_3 - Q_1}{2(P_{90} - P_{10})}\right)$	Sk $\left(S_k = \frac{Q_1 \cdot Q_3}{Md^2}\right)$	$CaCO_3$ %	Type of Sediment	Age
1.	154—157	0,0052	0,023	0,051	0,06	0,087	0,13	1,95	0,26	0,72	2,2	F. S.	Pleistocene
2.	157—160	0,004	0,06	0,30	0,37	0,44	0,52	2,71	0,38	0,29	3,2	M. S.	
3.	160—163	0,008	0,04	0,20	0,22	0,28	0,34	2,65	0,36	0,28	2,2	M. S.	
4.	163—166	0,014	0,11	0,25	0,28	0,30	0,34	1,65	0,29	0,52	3,5	M. S.	
5.	166—169	0,014	0,19	0,40	0,43	0,48	0,55	1,58	0,27	0,57	—	M. S.	
6.	169—171	0,17	0,27	0,38	0,40	0,48	0,55	1,33	0,28	0,89	4,4	M. S.	
7.	174—177	0,025	0,10	0,30	0,38	0,45	0,55	2,22	0,33	0,50	8,8	M. S.	
8.	177—180	0,001	0,0022	0,0047	0,0052	0,0085	0,018	1,96	0,18	0,85	5,3	CL.	Upper Pannonian
9.	185,5—188	—	0,00023	0,0011	0,0016	0,007	—	1,74	—	1,33	—	CL.	
10.	199,5—200,5	0,0032	0,015	0,053	0,052	0,10	0,14	2,58	0,31	0,53	3,5	C. Si.	
11.	202,2—202,8	0,0016	0,0044	0,012	0,013	0,028	0,05	2,52	0,24	0,85	4,4	F. Si.	
12.	204,8—205	0,041	0,09	0,10	0,10	0,12	0,17	1,15	0,37	1,08	3,5	S. S.	
13.	210,5—212,9	0,0046	0,0014	0,055	0,06	0,11	0,15	2,80	0,37	0,51	2,2	C. Si.	
14.	218,5—220,9	0,0035	0,010	0,027	0,034	0,06	0,095	2,45	0,27	0,82	—	C. Si.	
15.	225,5—226,9	0,035	0,065	0,096	0,011	0,11	0,15	1,30	0,19	0,76	8,8	S. S.	
16.	229,5—230,1	0,0015	0,0058	0,021	0,026	0,05	0,085	2,93	0,26	0,66	—	C. Si.	
17.	238,0—240,5	0,0042	0,011	0,033	0,036	0,065	0,094	2,43	0,30	0,66	4,4	C. Si.	
18.	240,5—242,9	0,0045	0,013	0,034	0,034	0,060	0,011	2,15	0,36	1,24	2,2	C. Si.	
19.	245,5—246,4	0,00081	0,0021	0,0052	0,0057	0,012	0,024	2,40	0,21	0,93	2,2	CL.	
20.	250,8—251,4	0,0017	0,0052	0,0090	0,0095	0,013	0,021	1,58	0,20	0,83	4,4	F. Si.	
21.	251,4—252,3	0,002	0,0051	0,0085	0,0095	0,013	0,019	1,60	0,23	0,92	4,4	F. Si.	
22.	254,0—255,3	0,0019	0,005	0,013	0,014	0,024	0,038	2,20	0,26	0,71	6,6	F. Si.	
23.	257,3—258,4	0,0007	0,0061	0,052	0,095	0,11	0,17	4,20	0,31	0,23	—	S. S.	
24.	262,4—263,2	0,0013	0,0026	0,0065	0,007	0,012	0,019	2,15	0,26	0,74	—	F. Si.	
25.	263,2—265,0	0,00055	0,0014	0,011	0,011	0,24	0,56	4,16	0,20	0,28	2,2	F. Si.	
26.	269,4—270	0,00095	0,0016	0,0031	0,004	0,0065	0,011	2,02	0,24	0,76	8,8	CL.	
27.	275,8—277	0,0008	0,0026	0,009	0,01	0,018	0,03	2,64	0,26	0,85	8,8	F. Si.	
28.	290,3—293,5	0,00052	0,0014	0,0031	0,0038	0,0067	0,011	2,19	0,25	0,97	2,2	CL.	
29.	295,0—295,4	0,0008	0,0017	0,0046	0,005	0,0097	0,018	2,39	0,23	0,78	5,7	CL.	
30.	295,4—297,4	0,0022	0,008	0,03	0,028	0,08	0,15	3,16	0,24	0,70	—	C. Si.	
31.	308,0—312	0,058	0,076	0,1	0,11	0,12	0,18	1,25	0,18	0,91	6,6	S. S.	
32.	312—317	0,065	0,12	0,2	0,21	0,3	0,4	1,58	0,27	0,90	3,5	M. S.	
33.	318,1—318,4	0,00095	0,0019	0,0044	0,0044	0,0076	0,018	2,00	0,17	0,75	—	CL.	
34.	332,9—333,5	0,00067	0,0017	0,005	0,0057	0,008	0,016	2,17	0,21	0,54	4,4	CL.	
35.	343,2—346,0	—	0,00065	0,0045	0,0055	0,009	0,013	1,17	—	0,29	0,9	CL.	
36.	357—358	—	0,0004	0,0052	0,0065	0,011	0,021	5,25	—	0,16	4,4	F. Si.	
37.	362—366	—	0,0013	0,006	0,0065	0,012	0,018	3,05	—	0,43	—	Ti. Si.	
38.	372—372,6	0,076	0,12	0,16	0,026	0,21	0,26	1,33	0,24	0,98	4,4	S. S.	
39.	378,5—382,2	0,03	0,051	0,08	0,084	0,11	0,15	1,47	0,25	0,88	6,6	F. S.	
40.	389,7—393,8	0,001	0,0034	0,009	0,009	0,014	0,021	2,03	0,26	0,59	12,6	F. Si.	
41.	398,2—402,0	0,00038	0,00075	0,0019	0,0023	0,004	0,0085	2,30	0,20	0,83	2,7	CL.	
42.	430,8—432,3	0,002	0,008	0,04	0,05	0,075	0,13	1,27	0,26	0,38	2,2	C. Si.	
43.	432,3—436,5	0,00046	0,0011	0,0031	0,0037	0,0067	0,012	2,47	0,24	0,77	2,2	CL.	
44.	447,2—454	0,021	0,05	0,1	0,11	0,15	0,19	1,74	0,29	0,75	3,5	S. S.	
45.	470,5—476	0,00038	0,0008	0,0024	0,003	0,0055	0,011	2,62	0,22	0,76	4,2	CL.	
46.	486,3—490	0,0038	0,016	0,07	0,08	0,11	0,13	2,62	0,37	0,36	6,6	F. S.	

CL. = CLAY < 0,005 mm ⌀, F. Si. = FINE SILT = 0,005 – 0,02 mm ⌀,
S. S. = SMALL SAND = 0,1 – 0,2 mm ⌀, M. S. = MEDIUM SAND = 0,2 – 0,5 mm ⌀
C. Si. = COARSE SILT = 0,02 – 0,06 mm ⌀, F. S. = FINE SAND = 0,06 – 0,1 mm ⌀,

values in the samples containing the greatest amount of basaltic hornblende (*Fig. 5*, IVth column).

Comparatively harder rocks cannot be encountered elsewhere, being limited to this depth. The cement is calcareous silica. The precipitation of this was promoted by the fact that with the decrease of the inland sea surface area, the waters of the influent rivers often changed the pH value of sea water. The large-scale growth of the seashore vegetation may have led to the same consequences.

Determined on the basis of changes in grain composition, the near-shore and farther off-shore phases have been indicated in stretches 2—3 of the IVth column of *Fig. 2* to *Fig. 5*. In this sector, in compliance with the regression-bound trend, the distance to the shoreline often changed in dependence on the changes of the marginal zone and of the equilibrium of accumulation and erosion.

b) In the 123-m-thick sequence between 177 and 300 m it is clay and silt that predominate. Accordingly, its lithological characteristics are also more balanced, so that the sequence is indicative of a slightly subsiding sedimentation basin. Occurring rather frequently, limonite nodules suggest a well-aerated lake-water environment of shallow depth (*Fig. 2* and *Fig. 3*). The heavy mineral composition and the feldspar ratio indicate the continuation of the conditions which characterized the upper part of the previous phase. Like in a number of other Great Plain Upper Pannonian profiles, the change of source area took place earlier than the inversion of regression into transgression [B. MOLNÁR, 1968*b*, 1968*c*]. Hence, the heavy mineral assemblage found at 204 m may represent the fore-runner of a final regression phase which is absent in the profile or which may even not have developed because of possible rapid emergence.

The Upper Pannonian sequence contains no consolidated rock within these uppermost 123 metres.

The entire Upper Pannonian sequence is devoid of fauna [M. ERDÉLYI, 1960]. However, M. FARAGÓ [1960] collected from the Upper Pannonian disclosed here, a pollen assemblage completely identical with that of the Upper Pannonian lignites occurring in the southern foreland of the Mátra Mountains. The Kemecse pollen assemblage also consists of shallow-water to paludal forms. This fact as well as the grading of the deeper, macrofossiliferous Upper Pannonian horizon into the pollen-bearing interval does not justify the assignment of the latter to the Upper Pliocene ("Levantine"). Neither does so the variegated, flamboyant nature of the colour of the sediments [M. SZÉLES, 1965], as F. BARTHA [in M. ERDÉLYI, 1960] described an Upper Pannonian fauna from a sedimentary sequence in the near-by Macs borehole, showing completely the same variegation *(Fig. 1)*.

2. *Pleistocene:* the 177-m-thick sequence can be split up into two parts:

a) The medium sands and fine silts of the 167—170 m interval are unconformable on the Upper Pannonian. The discrimination of this latter from the other Pleistocene developments is justified by that its source area is but partly identical with that of the others. East of the city of Nyíregyháza the morphology of the surface is particularly marked. The effect of the Upper Pannonian sediments removed from the higher portions of topography is reflected by the accumulation of the first Pleistocene strata of the deeper-seated Kemecse area.

b) The remaining 167 m of the Pleistocene shows a heavy mineral composition completely identical with, that produced by the present-day rivers of the north-eastern Great Hungarian Plain [B. MOLNÁR, 1964]. This sequence may have been deposited after the relatively late Pleistocene subsidence of the Nyírség area, recorded by J. SÜMEGHY [1944, 1955]. All these deformations are the result of a single, late crustal rhythm (subsidence), whereas in other areas, where the presence of older sediments can also be warranted, several sedimentation rhythmus can be shown to have occurred during the Pleistocene [B. MOLNÁR, 1967, 1968; A. RÓNAI, 1968].

Comparison of the Kemecse log with other near-by bore columns

The only 48-m-thick Pleistocene member of the Macs borehole drilled in the Hajdúság *(Fig. 1)* is substantially thinner than that of the Kemecse borehole. However, in its 453-m-thick Upper Pannonian section sedimentation rhythms, similar to those observed at Kemecse, could be revealed. In Pannonian time the Hajdúság was of higher hypsometric position than the Kemecse area was. Hence the frequency of lignite beds there. In the Upper Pannonian of Macs, four minor cycles have been revealed [B. MOLNÁR, 1965, 1966, 1968*b*]. However, the regression and transgression phases discovered at Kemecse are thicker than their two equivalents of Macs Accordingly, at Kemecse, where a sublittoral facies is present, the minor rhytms of crustal movements are not reflected perceptibly by sedimentation. At Macs the rhythms, which preceded the latest Pannonian final regression, were the shorter the closer to the completion of regression. The disappearance of the Pannonian inland sea was due to the gradual increase or the duration of uplift or static conditions opposed to the decrease of the duration of subsidence.

After the Upper Pliocene ("Levantine") break in sedimentation it is the Pleistocene that follows at Macs, too. However, it largely differs in lithology from the Pleistocene deposits of Kemecse. In Pleistocene time much of what is now the Hajdúság, inclusive of the Macs area, was a high-perched platform of Pannonian sediments, upon which there is no Pleistocene fluviatile accumulation. Despite its lower thickness, the Pleistocene sequence embraces a larger time span. Therefore, ten loess deposition phases, at least six soil genetic phases, and five wind-blown sand interlayers could be detected within the Macs profile.

CONCLUSION

In the northeastern Great Hungarian Plain the Upper Pannonian is represented by the following two types:

1. On the high-perched Pannonian platform of the Hajdúság there is a shallow-water, lacustrine, variegated sedimentary sequence with lignite seams which is in many respect similar to the contemporaneous developments of the Mátra— Bükkalja areas but which is constituted by sediments of increasingly smaller grain size farther off-shore. The source area seems to have been the marginal volcanic belt.

2. In the northern part of the Nyírség, in the vicinity of Kemecse, the lack of lignite is an evidence of a somewhat deeper, farther off-shore environment in which, however, variegated sediments like those of Macs were deposited, testifying to a shallow-water lake environment. Upwards in the column, the volcanic material is gradually outscored by the older sedimentary material of the marginal zone.

The lithology of the deposits of both the areas proves the rhythmicity of differential crustal movements. In the Hajdúság the single rhythms lasted for a shorter time, in the norther Nyírség for a longer time, so that the two areas also differ by the rate of accumulation of sediments within the span of time of the individual rhythms. Nota bene, in the Hajdúság the sediments accumulated within one cycle are thinner, in the northern Nyírség, thicker.

In the Hajdúság and the Nyírség the discrimination of the Upper Pliocene ("Levantine") is not justified by sedimentation. The author believes it to be absent there. The Hajdúság was the scenery of erosion in Latest Pliocene and Earliest Pleistocene time, whereas the Nyírség witnessed erosion even for a considerable part of the rest of the Pleistocene epoch.

The post-Pannonian red clay beds of the northern marginal zone are absent in many places of the basin's northeastern part. This seems to be due to break in sedimentation or to later erosion. The "red clay beds" of the Hajdúság cannot be identified with the red clays of the marginal zone. In fact, the former are loess-based soil layers of much coarser grain composition [B. MOLNÁR, 1966]. Where the red clays are absent, the Upper Pannonian — Pleistocene boundary can be traced on the basis of differences in lithology.

In the northeastern Great Hungarian Plain even the Hajdúság and Nyírség areas differ from each other in Pleistocene history.

The Hajdúság was abandoned by fluviatile accumulation in Pleistocene time and witnessed eolian accumulation processes for a considerable part of the Pleistocene.

In the Nyírség, in the early part of the Pleistocene, the higher-seated Pannonian ridges were eroded and the sediment worn away was deposited in the deeper-seated areas, giving rise to a thinner sequence (e.g. at Kemecse, too). That time the Nyírség rivers skirted the Hajdúság area and carried their load farther away, to deposit them in the depression of the Kőrös Riverine [J. URBANCSEK, 1965].

With the Late Pleistocene subsidence of the Nyírség [J. SÜMEGHY, 1955], the rate of fluviatile accumulation increased in the area. However, both loess and wind-blown sand are represented among the fluviatile sediments and above their sequence [Z. BORSY, 1961; A. RÓNAI—L. MOLDVAY, 1966]. As shown by recent data [GY. CSICSELY, 1968], in the western Nyírség the wind-blown sands play a considerable role even at greater depths than was demonstrated before.

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CLASSIFICATION OF MANGANESE DEPOSITS

SUPRIYA ROY

ABSTRACT

The widely distributed manganese deposits of the world had earlier been classified genetically by PARK [1956]. VARENTSOV [1964] classified principal manganese formations of exogenetic type on the basis of paragenetic associations of rocks. Both the classifications have their limitations. An attempt has been made in this paper to present a genetic-associational classification of manganese deposits. Accordingly the principal manganese deposits of the world can be classified into three broad genetic types e.g. hydrothermal, sedimentary and superficial. The sedimentary type has been genetically subdivided into nonvolcanogenic and volcanogenic types depending upon their source of the metal. Both the nonvolcanogenic and volcanogenic deposits have further been subdivided according to characteristic rock associations. It has been shown, however, that no generalized conclusion can be drawn to relate the associational subdivisions to particular genetic types or any unique tectonic set up, and thus the associational subdivisions have only a descriptive value.

INTRODUCTION

Manganese ore deposits are widely distributed in the continents and on ocean floors all over the world. Even a very conservative estimate indicates the reserve of manganese ores of the world to well over a billion tons, taking into consideration the well known deposits on the continents only. VARENTSOV [1964] estimated that more than 70% of the total manganese deposits of the world are of Cenozoic age. The Mesozoic era is conspicuous for the paucity of manganese ore deposits (only about 0,004% of the world reserve) while the Paleozoic and Precambrian eras share the rest of the known deposits almost equally.

The above estimate does not include the recent deposits of manganese nodules on ocean floors, the potentiality of which has been emphasized by a number of workers including MERO [1965], BONATTI and NAYUDU [1965], STRAKHOV [1966], PRICE [1967], BÖSTROM [1967] and others.

These vast deposits of manganese have originated by diverse processes and are associated with diverse rock types. Systematic attempts of genetic and/or associational classifications are few and are mostly incomprehensive. The author will attempt here to discuss these earlier classifications and put forward a new scheme that may bring into purview most of the well known deposits of the world.

EXISTING CLASSIFICATIONS

PARK [1956] suggested a genetic classification of manganese ores. His scheme is given below in Table 1.

TABLE 1

Genetic classification of manganese ores [PARK, 1956]

MANGANESE ORES

- HYDROTHERMAL DEPOSITS
- SEDIMENTARY DEPOSITS
  - DEPOSITS INDEPENDENT OF VOLCANIC ACTIVITY
  - DEPOSITS ASSOCIATED WITH TUFFS AND CLASTIC SEDIMENTS OF VOLCANIC AFFILIATION
  - DEPOSITS ASSOCIATED WITH IRON FORMATIONS
- LOW TEMPERATURE SILICATE AND HAUSMANNITE DEPOSITS ASSOCIATED WITH SUBMARINE FLOWS
- METAMORPHIC DEPOSITS
- RESIDUAL ACCUMULATIONS LATERITES

This is the first systematic attempt to classify manganese deposits of all genetic types. However, some of the subdivisions proposed by PARK are not entirely acceptable in purely genetic scheme. As for example, within the "Sedimentary Deposits" PARK suggested three subdivisions on the basis of rock association. In a purely genetic classification of this type [cf. PARK, 1956, p. 75] such associational subdivisions should be kept separate to avoid confusion. Attention may particularly be drawn to the two subdivisions, "Deposits independent of volcanic activity" and "Deposits associated with iron formation", both of which are non-volcanogenic chemical sediments. Thus they should not be classed separately in a genetic scheme. Rather, they may be shown as different associational types under the broader genetic class of non-volcanogenic sediments.

Another subdivision, "Deposits associated with submarine flows and composed mainly of low temperature silicates and hausmannite", proposed by PARK, seems to be superfluous. PARK included "Deposits of complex manganese silicates and oxides" which are associated with submarine pillow flows, under this class. He mentioned the Olympic Peninsula deposit, Washington, Franciscan formation, California and the Japanese deposits as examples. Of the above, the Olympic Peninsula deposit and those of the Franciscan formation can be classed appropriately with the volcanogenic-sedimentary deposits. The Japanese deposits may be included under the hydrothermal [LEE, 1955] and/or volcanogenic-sedimentary class which were later thermally metamorphosed [WATANABE, 1959; 1960; WATANABE, KATO and ITO, 1960].

The class, "Metamorphic Deposits" suggested by PARK is also superfluous as a genetic type, as in none of the deposits cited by him (Indian and Brazilian deposits) any effective concentration of manganese took place during metamorphism. In all these cases, only pre-existing sedimentary manganese deposits were later metamorphosed resulting in compaction and reconstitution of mineral phases, but not necessarily producing or improving the quality of the ores. Thus, these orebodies should appropriately be classed under sedimentary deposits, the metamorphism being only a superimposed phenomenon. Moreover, the manganese oxide ore deposits of Brazil are, mostly (with the exception of those associated with itabirites) supergene products formed by oxidation of meta-sedimentary manganese carbonate protore.

VARENTSOV [1964] attempted to classify the principal manganiferous formations as paragenetic associations of rocks in which manganese deposits are characteristic. His classification is given in Table 2.

TABLE 2

Classification of manganese formations as paragenetic association of rocks [VARENTSOV, 1964]

MANGANESE FORMATIONS

NIKOPOL MANGANIFEROUS FORMATION (Orthoquartzite Glauconite-clay)	THE LIMESTONE DOLOMITE GROUP	GONDITE FORMATION	SEDIMENTARY VOLCANIC FORMATIONS OF THE GREENSTONE SERIES	GROUP OF SUPERIMPOSED CHIEFLY LATERITIC FORMATIONS

THE JASPILITE GROUP	SILICEOUS-SHALE-ORTHOQUARTZITE GROUP	SEDIMENTARY-VOLCANIC FORMATIONS OF THE PORPHYRY SERIES	FLYSCH (TUFFACEOUS TERRIGENOUS) FORMATIONS

In line with the objection raised on assigning the "Metamorphic Deposits" a separate class in PARK'S classification, the author considers that in VARENTSOV'S classification also, the type "Gondite Formation" is superfluous. The type "Flysch Formation" has also been objected to by DZOTSENIDZE [1966] who expressed doubts about their precise nature. In any case, even according to VARENTSOV, the Flysch type manganese formations are extremely rare. VARENTSOV'S classification is also limited in scope as he attempted to classify the manganese formations of exogenic type only and has not dealt with the hydrothermal deposits.

SHATSKIY [1964] considered most of the manganese ore deposits of the world to be volcanogenic and suggested that the ore deposits of this type can broadly be classified into two types: Greenstone-Siliceous Group and Porphyry-Siliceous Group. The details of these two types will be discussed at length later in the paper.

GENETIC TYPES OF MANGANESE FORMATIONS

In this text, the author will follow a broad based, three fold genetic classification of manganese formations of the world, viz. *1.* Deposits formed by hydrothermal process, *2.* Deposits formed by sedimentary process, and *3.* Deposits formed by superficial concentration. These types will be briefly discussed below:

1. Deposits Formed by Hydrothermal Process

Hypogene vein deposits of manganese formed by hydrothermal process have been described from different parts of the world [HARIYA, 1961; HEWETT, 1964; HEWETT and FLEISCHER and CONKLIN, 1963, etc.]. A close genetic correlation of these hypogene veins and recent hot spring apron deposits of U.S.A. and Japan, has been established by the above workers. Recent concentrations of manganese nodules on ocean floors have been suggested to have formed from hydrothermal solutions by VON GUMBEL [1878], WEDEPHOL [1960], CRONAN and TOOMS [1967] and others.

HEWETT and FLEISCHER [1960] showed that in hypogene veins in different parts of the U.S.A., particularly in San Juan region of Southwest Colorado, rhodonite and rhodochrosite follow the deposition of common sulfides of copper, lead, zinc, silver and the less common sulfoarsenides and sulfoantimonides of copper and silver with quartz and barite. They also showed that in many instances, hypogene veins composed of manganese minerals also contain barite, fluorite and huebnerite. Such association of hypogene manganese minerals in veins with barite, fluorite, gold-silver and base metals, has been elaborated by HEWETT [1964]. HEWETT showed that in zoned hypogene veins, manganese is present in manganous state in the minerals rhodochrosite, rhodonite, alabandite and huebnerite in the lowest two zones, in association with base metals and gold-silver minerals, respectively. Manganese is present in a more oxidized state in the higher zones, as oxides in black calcites associated with fluorite and barite and finally as higher oxides of manganese (pyrolusite, cryptomelane, psilomelane, hollandite, manganite) in the uppermost zones.

HEWETT and FLEISCHER [1960] and HARIYA [1961] have demonstrated that the present day hot springs are depositing manganese oxides (cf. Hot Spring No 23, Arkansas; Saline Valley, California; Sodaville, Mineral County, Nevada; Komaga-Dake Cold Spring, Iwao hot spring, Niimi hot spring and Akan hot spring, Japan: Table 3, ROY, 1968) and thus testify to the feasibility of formation of manganese deposits from a hydrothermal source. NIINO [1959] also demonstrated that sea-floor springs off the southeast coast of Japan, empty manganese-rich solutions in the ocean.

In the ancient deposits, paragenetic association of the mineral groups of manganese, barite, fluorite, gold-silver and base metals and the zoning exhibited by them, confirm the hypogene nature of the manganese minerals, formed from ascending hydrothermal solution (Examples: Pioche, Eureka county, Nevada; Leadville, Pitkin County, Colorado; Bisbee, Tombstone, Gila, Graham, Greenlee and Pinal County, Arizona; Silver City, Hidalgo, Luna, Socorro and Dona Ana County, New Mexico; Philipsburg and Butte, Montana; Inakuraishe, Hokkaido, Japan; Dzhedza and Nayzatas, Central Kazakhstan, U.S.S.R. etc.).

HEWETT [1966] proposed that for most of the stratified deposits of manganese containing little or no iron, the metal was supplied by veins and aprons produced by thermal waters with possible volcanic affiliation. He thought that the absence of commensurate iron with the manganese deposits in space and time can only be explained by postulating a source from ascending hot springs, where iron is separated from manganese by precipitating in deeper zones. Thus HEWETT postulated hydrothermal solutions as the ultimate source of manganese and, in his opinion, the veins and aprons formed by thermal waters, after reworking, gave rise to stratified deposits by remobilisation of the metal.

The above contention of HEWETT [1966] is based on the assumption that iron is separated from manganese at depth during precipitation from thermal waters and hence this process alone can explain the lack of commensurate iron in many stratified deposits. The separation of iron from manganese in vertical columns due to differential mobility during diagenesis of sediments has recently been emphasized by LYNN and BONATTI [1965], STRAKHOV [1966] and BÖSTROM [1967] and this concept certainly restricts HEWETT'S hypothesis from universal application [cf. ROY, 1968].

Though the extent of the role of hydrothermal process in the formation of manganese deposits remains controversial, it is evident that deposits formed by this process constitute a recognised genetic type. Deposits of unequivocal hydrothermal origin, however, seems to be few and possibly account for only a minor part of the manganese deposits of the world [BÖSTROM, 1967].

2. Deposits Formed By Sedimentary Processes

The deposits formed by sedimentary processes constitute, by far, the majority of the commercial deposits of manganese of the world. Genetically, the sedimentary manganese formations can be subdivided into two broad types: *(a)* Volcanogenic-sedimentary deposits where the metal was supplied from a volcanic source and the precipitation was closely related in time and space to subaqueous volcanic eruptions (exhalative-sedimentary); and *(b)* non-volcanogenic sedimentary deposits where the source and precipitation of manganese was not connected with any volcanic episode and the metal was entirely derived by weathering of continental landmass (pure sedimentary).

The proponents of volcanogenic-sedimentary manganese formation include BOULADON and JOURAVSKY [Morocco deposits; 1952, 1956], GEIJER and MAGNUSSON [Swedish deposits; 1948], ÖDMAN (Swedish deposits; 1950], PARK [Olympic Peninsula deposit, U.S.A. 1946], SERVICE [Nsuta deposit, Ghana; 1943], SUSLOV [Kuznetsky Altai deposit; 1967) and others, who suggested that concentration of manganese in these deposits took place either during direct volcanic activity or by weathering of manganese-bearing volcanic rocks. SHATSKIY [1964] elaborated this concept and considered that most of the important sedimentary manganese deposits of the world are of volcanogenic derivation, excepting only some of those associated with iron formations (cf. Morro do Urucum, Brazil).

The mechanism of formation of the exhalative-sedimentary type of manganese deposits by leaching out of metals from contemporaneous subaqueous eruptions, has been explained by PARK [1946], KRAUSKOPF [1956] and others. This hypothesis of concentration of manganese by leaching from subaqueous volcanic eruptions (hyaloclastites), has been considered to be the operative process, for the formation of recent deep-sea manganese nodules by BONATTI and NAYUDU [1965], BONATTI [1967], HEWETT, FLEISCHER and CONKLIN [1963], SUMMERHAYES [1967] and others.

A number of workers [BETEKHTIN, 1937; DORR *et al*, 1956; ROY, 1966; STRAKHOV, 1966; STRAKHOV and SHTERENBERG, 1966; VARENTSOV, 1964 etc] proposed that many of the important manganese deposits are of "pure sedimentary" type i.e. nonvolcanogenic in source (cf. Chiatura, Nikopol, Bolsh'e Tokmaksk, Labinsk, Maliy Khingan, Usinsk, U.S.S.R; Minas Gerais, Bahia, Morro do Urucum, Matto Grosso, Brazil; Madhya Pradesh — Maharashtra, Gangpur, Srikakulam, India and others). These deposits have been conclusively proved to be unconnected with any volcanic episode. The manganese in these deposits was evidently derived by weathering of continental rocks. A similar conclusion about the formation of deep-sea manganese nodules has been drawn by GOLDBERG [1954] and GOLDBERG and ARRHENIUS [1958].

The evidences suggesting a volcanogenic derivation of manganese for sedimentary deposits are as follows: *(i)* spatial contiguity of volcanic rocks with manganese formation *(ii)* field features of the manganese deposits themselves, such as interlayering and interfingering of manganese formations and volcanic rocks, combination of vein and stratified deposits, association with hydrothermal deposits etc., *(iii)* hypogene alteration of associated rocks and co-precipitation of chemogenic rocks directly related to volcanic activity, and *(iv)* higher content of minor elements.

None of the above evidences, however, is considered to be unequivocal by itself. Thus, the manganese formations associated with volcanic rocks in S. E. Newfoundland [MOHR, 1965], Central Poland [SAMSONOWICZ, 1956] and Usinsk, U.S.S.R. [VARENTSOV, 1964] have been conclusively proved to be nonvolcanogenic. The use of high concentration of minor metals as an evidence for volcanogenic source also has been questioned by STRAKHOV [1966] who suggested that the higher content of minor elements in some manganese deposits may not be the effect of volcanic activity alone but might have been subscribed by both terrigenous and volcanic sources.

Thus it is clear that the occurrence of unequivocal volcanogenic-sedimentary deposits of manganese is not as widespread as it was thought to be and a similar conclusion has been drawn by BÖSTROM [1967] in case of deep-sea manganese concentrations. BÖSTROM has shown that the role of submarine volcanism is relatively minor in the formation of nodules on ocean floors and stated "at 100% leaching efficiency, only 5% of the total excess manganese could be derived from the effused volume of basalts in the Pacific".

The general consensus among most of the workers is that both volcanogenic and non-volcanogenic sedimentary manganese deposits are common [cf. VARENTSOV, 1964; STRAKHOV, 166 etc.]. The dual source of manganese for the oceanic nodules also has been suggested by ARRHENIUS, MERO and KORKISCH [1964], KRAUSKOPF [1967], SKORNYAKOVA, ANDRUSHCHENKO and FOMINA [1962], STRAKHOV [1966] and others.

In any genetic classification of manganese formations, therefore, both volcanogenic and nonvolcanogenic sedimentary deposits should find adequate places, though, admittedly, there may be such transitional cases where the two types can hardly be distinguished.

The effect of diagenetic modification of the manganese sediments has been emphasized by LYNN and BONATTI [1965], STRAKHOV [1966], STRAKHOV and SHTERENBERG [1966] and others and similar ideas have been forwarded in case of iron sediments by LEPP [1968]. It is difficult to determine, at this stage, how far the mineralogy of the sedimentary manganese ores had been controlled by conditions of precipitation or diagenetic processes. In some cases, the sediments have later been subjected to regional or contact metamorphism and thoroughly modified (cf. India, Brazil, Ghana etc.).

3. Deposits Formed By Superficial Concentration

Deposits of manganese oxide, formed by supergene agencies at or near surface, are common in different parts of the world and some of them assume considerable dimensions to be regarded as major commercial deposits. Concentration of manganese oxides in these deposits is effected by alteration and remobilization of either earlier manganese formations or from rocks that initially contained manganese as minor constituent.

The formation of manganese oxides by alteration and remobilisation of earlier formations where manganese was a major constituent, is primarily restricted to the change in oxidation state of the manganese and thereby formation of new phases in low temperature-pressure conditions. The trend of such changes by oxidation and hydration has been shown to be dependent on the source rock and the oxidation gradient [BRICKER, 1965, ROY, 1968]. Thus a manganese carbonate protore (with primarily rhodochrosite) should ultimately be oxidised to pyrolusite (β-MnO_2) and/or cryptomelane (α-MnO_2) depending upon the extent of K-absorption form ground water [cf. Minas Gerais, Brazil, HOREN, 1953, MARVIN and ZWICKER in BRICKER, 1965; Moanda, Gabon, Africa, BAUD, 1956; Philipsburg, Montana, U.S.A., LARSON, 1962, PRINZ, 1967; Butte, Montana, U.S.A., ALLSMAN, 1956, FLEISCHER, RICHMOND and EVANS, 1962; Piedras Negras, Mexico, ZWICKER in BRICKER, 1965; Ghana, Africa, SOREM and CAMERON, 1960, ZWICKER in BRICKER, 1965; Toyoguchi Mine, Iwate Prefecture, Japan, NAMBU and TANIDA, 1961; Úrkút, Hungary, CSEH NEMETH and GRASSELLY, 1966; etc.]. Depending upon, the oxidation gradient, however, such alteration of rhodochrosite may be arrested in intermediate stages giving rise to γ-MnO_2 (nsutite) or δ-MnO_2 (birnessite). A manganese silicate protore (such as gondite), on the other hand, does not yield γ-or δ-MnO_2 at any stage of alteration, even in ideally low oxidation gradient and directly changes over to pyrolusite or cryptomelane. Pre-existing metamorphosed lower oxide ores also show considerable alteration due to oxidation by supergene agencies, but they (chiefly composed of braunite, bixbyite, jacobsite, hausmannite etc.) also change directly to pyrolusite and/or cryptomelane.

Superficial concentration of manganese from country rock containing only a small amount of the metal has been reported from many countries. Limestones and dolomites are well known in this regard and they contain enormous quantity of manganese locked in them, though the distribution is very sparse and the percentage of a metal rarely exceeds 2—3%. Shales, phyllites and quartzites also contain manganese in very low concentration. Thus in U.S.A. in the southeastern states extending from Pennsylvania through Maryland and Virginia to Georgia, Alabama and Arkansas, superficial manganese deposits have formed in residual clays overlying carbonate and other rocks of Paleozoic rocks that originally contained some amount of manganese. In the Piedmont mine, Cambell county, Central Virginia, almost half the thickness of limestone contains from 0.50 to 0.75% manganese and this is considered to be the source of the superficial manganese oxide deposits that are associated with the limestone [HEWETT and FLEISCHER, 1960, p. 16]. Similarly the shaly dolomite from Shady Valley, and Bumpass Cove, Tennessee contain 0.39 to 1.24% and 0.13 to 0.83% manganese respectively, which is responsible for the formation of the associated superficial deposits. Superficial deposits of manganese are also found in cherts and quartzite (Weissner quartzite and Fort Payne Chert, U.S.A.) as also in phyllites and shales (Orissa and Mysore deposits, India) and at present the source of the metal is assumed to be the country rocks.

SUBDIVISIONS OF THE GENETIC TYPES BASED ON ROCK ASSOCIATIONS

Association with particular rocks is not very characteristic with both epigenetic hydrothermal deposits and superficial deposits of manganese. In case of sedimentary deposits, however, association of characteristic rock types sometimes assumes considerable importance. Several workers have tried to interpret such associations in terms of environments during deposition. VARENTSOV [1964] first classified the sedimentary manganese deposits according to rock associations (see Table 2).

A. Sedimentary Manganese Deposits of Nonvolcanogenic Source

Nonvolcanogenic, pure sedimentary deposits of manganese are found to be associated with the following principal rock types:

1. Association with Orthoquartzite — Glauconite — Clay and Orthoquartzite — Carbonate formations.

Vast deposits of syngenetic manganese ores of Cenozoic age, either unmetamorphosed or slightly metamorphosed, are associated with orthoquartzite — glauconite — clay formations at Chiatura, Nikopol, Bols'he Tokmaksk, Labinsk, Mangyshlak and other deposits of U.S.S.R. and Timna Dome, Israel. Generally nonvolcanogenic is envisaged [BETEKHTIN, 1936, 1937, SOKOLOVA, 1964, VARENTSOV, 1964 and others] though DZOTSENIDZE [1966] concluded that the Chiatura deposit is of "remote volcanogenic type" [cf. SHATSKIY 1964]. STRAKHOV and SHTERENBERG [1966] however, proved conclusively that the arguments of DZOTSENIDZE are untenable. The deposits occurring in association with orthoquartzite-glauconite-clay formation are generally developed on a stable platforms or areas close to platforms in stable areas of the crust. The manganese ores generally consist of higher oxides (pyrolusite, cryptomelane etc.) which pass through a mixed type rich in manganite, to manganese carbonate (rhodochrosite). This variation of mineralogy had been explained by BETEKHTIN [1936, 1937] to be due to depositional condition. VARENTSOV [1964], STRAKHOV [1966] and STRAKHOV and SHTERENBERG [1966], however, explained this gradual change in mineralogy from higher oxides to carbonates through manganite, as due to effects of diagenesis. The high terrigenous impurity of the ore is reflected in the high content of SiO_2. The manganese formations of this associational type apparently originated as a result of severe weathering of continental rocks and later deposition of the ore in shallow littoral areas of marine basins or lagoons. The Nikopol and Chiatura deposits were formed in a humid, the Mangyshlak in semi-arid and the Timna Dome deposit, Israel, in arid condition [VARENTSOV, 1964].

The regionally metamorphosed manganese formations of Sausar and Gangpur Groups of Precambrian age in India occur as part of the orthoquartzite-carbonate formation of possibly miogeosynclinal type [NARAYANSWAMI *et al*, 1963]. These have been conclusively proved to be of nonvolcanogenic sedimentary type [ROY, 1966]. The deposits are characterized by high temperature lower oxide assemblages (braunite-bixbyite-jacobsite-hausmannite) in the ores and manganese silicates (spessartite-quartz and manganese amphiboles and pyroxenes) in the associated gondites [ROY, 1966]. ROY and MITRA [1964] ROY [1966] and ROY and PURKAIT [1968] have shown that neither manganese carbonate nor low temperature silicate was present in the original sediments and the entire manganese was deposited as oxides or hydroxides. The lower oxides and silicates now constituting the ores and the gondites respectively are products of transformation and reaction during regional metamorphism.

2. *Association with Iron Formation*

The universal presence of undifferentiated manganese in varying quantity has been reported from different facies of sedimentary iron formation by JAMES [1954, 1966] and LEPP [1963, 1968]. Concentration of manganese as important ore deposits, characteristically associated with iron formation, have also been described from different countries including Minas Gerais, Bahia and Morro do Urucum (Brazil), Postmasburg and Kalahari (Africa), Maliy Khingan (U.S.S.R) etc. These deposits are unequivocally considered to be nonvolcanogenic by most of the previous workers including PARK [1956] and even by such staunch supporters of volcanogenic origin of manganese deposits as SHATSKIY [1964]. VARENTSOV [1964] has shown that these manganese deposits are situated in different tectonic set up. The eugeosynclinal type is represented by Minas Gerais and Bahia deposits (Brazil) and Postmasburg — Kalahari deposits (Africa). The miogeosynclinal and platform types are represented by Maliy Khingan (U.S.S.R.) and Morro do Urucum (Brazil) deposits respectively.

In Minas Gerais, Brazil, meta-sedimentary manganiferous formations occur in three associations [DORR *et al.*, 1956]:

(i) Manganese silicate-carbonate-sulphide protore.
(ii) Marble-itabirite protore where manganese was deposited as part of the chemical sediments.
(iii) The clastic sediments now represented by phyllite, quartzite etc.

The manganese silicate-carbonate-sulphide protore is represented by rhodochrosite-manganoan calcite-alabandite-spessartite-rhodonite-manganoan cummingtonite-thulite-tephroite-pyroxmangite-neotocite-bementite-graphite assemblage.

DORR *et al* [1956] concluded that the manganese formation was originally syngenetically deposited in negative Eh and pH around 7 in an euxinic environment, resulting in mineral assemblages of manganese carbonate and sulphide. These were later regionally metamorphosed to give rise to the manganese silicate-carbonate-sulphide protore. Any concentration of manganese oxides in these rocks was due to supergene agencies.

In marble-itabirite protore of Minas Gerais, Brazil, manganese is concentrated in both the members independently. The manganese content in itabirite varies from 0,1 to 45% Mn [DORR *et al* 1956]. The distribution of manganiferous itabirite in normal nonmanganiferous member is strictly stratigraphic. The manganiferous itabirite is, in all probability, primary in origin and does not show any indication of previous deposition as carbonates [PARK, DORR, GUILD and BARBOSA, 1951]. In marbles, on the other hand, manganese is considered to be locked up in manganoan calcite and dolomite and up to 4,20% MnO has been reported in the rock. On decomposition due to weathering, this manganese has been concentrated to form oxide orebodies of local importance.

The Postmasburg-Kalahari deposit of the Union of South Africa, considered to be part of a mobile belt that extends northwards into the Kalahari from the Orange river near Prieska [De VILLERS, 1956], provide another example of close association of manganese with banded iron formation. These is, however, considerable controversy about the origin of these deposits. SCHNEIDERHÖHN [1931] considered these deposits to be meta-sedimentary. J. E. DE VILLIERS [1944] concluded that the deposits are hydrothermal in origin, while the more recent workers [cf. J. DE VILLIERS, 1956] agree that the ores have formed by supergene concentration. The mineralogy of the ores (braunite-bixbyite-hausmannite-jacobsite), however, clearly indicates a high temperature origin. The laminated nature of braunite and hausmannite ore, conformable to the banded iron formation in the Smartt area in particular [J. DE VILLIERS, 1956] indicates a syngenetic formation. Thus, the possibility that the manganese formations were sedimentary in origin and later modified by metamorphism, cannot be ruled out. Such syngenetic meta-sedimentary ore deposits have been described from Otjosondu area in Damara System of southwest Africa [ROPER, 1956], where DE VILLIERS [1951] studied the mineralogy in detail. The mineralogy of the ores of Otjosondu and Postmasburg is comparable, though there are minor differences in textural details.

Syngenetic meta-sedimentary manganese deposits are closely associated with iron formations in a a miogeosynclinal sequence of Maliy Khingan area, U.S.S.R. [ILLARINOVA, KAMINSKAYA and NEMRYUK, 1958 cited by VARENTSOV, 1964; CHEBOTAREV, 1960]. The regionally metamorphosed manganese formation is constituted of the following mineral assemblage: braunite, hausmannite, hematite, magnetite, rhodonite, bustamite, tephroite, rhodochrosite, tremolite, actinolite, chlorite, sericite etc. The mineralogy indicates that the manganese was originally deposited in the sediments as oxides and carbonates. CHEBOTAREV [1960] compared these deposits with those at Morro do Urucum, Matto Grosso, Brazil.

Manganese oxide deposits associated with iron formation of platform type have been described from Morro do Urucum, Matto Grosso, Brazil [PARK *et al*, 1951; BARBOSA, 1956; VARENTSOV, 1964; SHATSKIY, 1964]. Here, manganese ores are interbedded with banded iron formation as part of the Banda Alta formation of Jacadigo Series. The ores are composed of higher oxides (mainly cryptomelane) and hydroxides of manganese [BARBOSA, 1956]. It is indicated that the manganese is genetically related to iron and silica. PARK *et al* [1951] showed that above and below each bed and lens of manganese oxide, beds of clastic materials (arkose) are found which suggest an abrupt but very temporary change from chemical to clastic sedimentation. Possibly this temporary change in environment somehow inhibited the precipitation of iron and silica and at the same time encouraged the precipitation of manganese (possibly by incursion of fresh water).

3. Association of Limestone — Dolomite Formation

Association of sedimentary manganese orebodies with limestone-dolomite formation is not uncommon. Such deposits are found in volcanic associations (cf. Morocco) and they may also be essentially nonvolcanogenic in derivation. An example of nonvolcanogenic manganese deposit forming part and parcel of an extensive limestone-dolomite sequence, has been presented by VARENTSOV [1964] at Usinsk, U.S.S.R. (Lower Cambrian). At Usinsk manganese carbonates are associated with limestone-dolomite formation in an eugeosynclinal tectonic set up. The deposits are made up of rhodochrosite, ferroan rhodochrosite, manganoan calcite and manganoan dolomite. The upper Permian deposit of Ulu-Telyaksk (W. Ural, U.S.S.R.) is a stable platform type manganiferous limestone-dolomite formation. The ore deposits are characterized by manganiferous limestones locally enriched to higher oxides by oxidation [BETEKHTIN, 1946].

B. Sedimentary Manganese Deposits of Volcanogenic Source

STRAKHOV [1967] distinguished three lithologic types among volcanogenic-sedimentary formations, viz. *(i)* the volcanic formations proper, including lavas and tuffs with no marked admixture of terrigenous material, and distinctive of ordinary platform segments of the earth: *(ii)* the volcanic-terrigenous formation with lavas, tuffs and sandstone and clay, formed chiefly in the sea; and *(iii)* the volcanic-siliceous formations with lava, tuff, terrigenous rocks and jasper and siliceous shales, developed in central parts of geosynclinal zone. Deposits of volcanogenic manganese, according to STRAKHOV, are associated with the third type of formation.

SHATSKIY [1964] also agreed that most of the known deposits of manganese ores of volcanic-sedimentary type are paragenetically associated with volcanogenic-siliceous facies. As already stated, SHATSKIY showed that the manganese deposits associated with volcanogenic-siliceous facies can broadly be classified into two subdivisions e.g. Greenstone-siliceous group and Porphyry siliceous group, according to the type of the parent volcanic rocks. These two subdivisions can further be classified into different lithologic formations as follows [SHATSKIY, 1964]:

I. GREENSTONE – SILICEOUS GROUP

GREENSTONE FORMATION	JASPEROID FORMATION	SILICEOUS-SHALY FORMATION	REMOTE SILICEOUS FORMATION

II. PORPHYRY – SILICEOUS GROUP

PORPHYRITIC FORMATION	SILICEOUS-SHALY FORMATION	REMOTE SILICEOUS FORMATION

According to SHATSKIY, these groups (excepting the Remote Siliceous formation in both cases) individually form single genetic series and any of the formations grade into others in the field. He, however, put up an word of caution in recognizing 'Remote Siliceous formation' in either of the groups. He observed: "Connections of the Remote Siliceous formations with volcanic ores can only be indirectly ascertained" and " Identification of isolated *Remote Siliceous Formations* within sedimentary series is a very difficult task. Only those of the formations, whose membership in the volcanogenic-siliceous series could be proved, should be assigned to this class". SHATSKIY also admitted that, "In the Remote Siliceous Formation (Porphyry-Siliceous Group) manganese ores, even if they are formed, are scarce and poor". The identity of Remote Siliceous formations has also been challenged by other workers including STRAKHOV and SHTERENBERG [1966]. It is, therefore, evident, in the light of the uncertainties pointed out above, that the Remote Siliceous formation in both the Greenstone-Siliceous Group and the Porphyry-Siliceous Group, is not an well established genetic or associational type and is, therefore, to be treated with caution.

In the Greenstone-Siliceous Group, the Greenstone formation is characterized by spilite, keratophyre, diabase and such other basic volcanic rocks. Volcanogenic manganese deposits in such association have been reported from South Ural, U.S.S.R. [SHATSKIY, 1964], Olympic Peninsula, U.S.A. [PARK, 1946], Oriente Province, Cuba [PARK *et al* 1944; SIMONS & STRACZEK, 1958] the western Alpine and Penine ophiolitic zones in Switzerland and Italy [GEIGER, 1948] Srednegorsk, Pozharevo area, Bulgaria [KOSTOV, 1944; SUSLOV, 1967], and others. The Jasperiod formation is characterized by jasper, tuff, subordinate limestone lenses and locally terrigenous rocks and it merges to Greenstone formation or Siliceous — Shaly formation by facies gradation. The important examples of manganese deposits associated with this formation are: the Parsetten and Faletta deposits of Graubünden Canton, Switzerland, Chevlyanovich deposit, Bosnia, Balkans and the deposits in the Franciscan formation, California, U.S.A. The deposits at Graubünden Canton, Switzerland are interbedded with Upper Jurassic radiolarites with layers of clayey shales, and this Jasperoid formation is underlain by ophiolitic greenstone formation. The manganese ores consist of oxides and carbonates. At the Chevlyanovich deposit also, the ore deposits are interbedded with Jurassic radiolarites. In this deposit braunite is the chief ore mineral [GEIGER, 1948]. In the Franciscan formation, California, U.S.A, manganese deposits are interbedded with radiolarian jaspers which, associated with carbonate rocks, form the top of the formation consisting of arkose, argillite and spilite-keratophyric basic intrusives. The deposits are characterized by manganese carbonate minerals. [TRASK *et al*, 1950].

Deposits of manganese in Siliceous-Shale formation of Greenstone-Siliceous Group, are comparatively rare. Important examples of deposits of this type are Kellerwald and Harz mountain deposits (Elbingerode and Lautenthal), Germany, Huelva Province, Spain, Fortuna Harbour, N. Newfoundland, Machang, Satakhun and Trenggan, Molucca Peninsula, Mazul'skoye deposit, U.S.S.R. and the Nsuta deposit, Birrim System, Ghana. All these deposits are enclosed in siliceous shales or their metamorphosed equivalents which are directly related to Jasperoid or Greenstone formation of Greenstone-Siliceous group.

The Porphyritic formation, represented by such volcanic rocks as quartz porphyry, dacite, rhyolite etc., contain well developed manganese oxide ore deposits, and in contrast to that of the Greenstone-Siliceous series, manganese is more concentrated as ore bodies and less dispersed in other rocks in the Porphyritic formation. The Glib-en-Nam deposit (Morocco) and Kolningsberg and Långban deposits

(Sweden) occur in Porphyritic formation. The Siliceous-Shaly formation in Porphyry-Siliceous Group is represented by the Central Kazakhstan deposits, U.S.S.R. [SHATSKIY, 1964].

SUGGESTED SCHEME OF CLASSIFICATION

The author has already attempted to review critically the earlier classifications of manganese formations and has discussed the broader aspects of the mode of origin of the different manganese formations and the rock types in which they occur. It has been shown that the five-fold genetic classification of manganese deposits [PARK, 1956] can be streamlined and made more broad-based by accepting a three-fold scheme (e.g. hydrothermal, sedimentary and superficial types). In the light of discussions already made on the source of manganese in the sedimentary deposits, the latter should be genetically subdivided into non-volcanogenic and volcanogenic types.

The sedimentary manganese deposits throughout the world have been shown to be characteristically associated with certain rock formations. No characteristic genetic implication of such association could, however, be drawn in all cases. The non-volcanogenic sedimentary manganese deposits occur in either of the three rock associations, viz. orthoquartzite-glauconite-clay and orthoquartzite-carbonate formations, iron formation, limestone-dolomite formation. These rock associations, but for the orthoquartzite-glauconite-clay formation [of Nikopol type; VARENTSOV, 1964] may be either geosynclinal or platform type. Thus, manganese deposits associated with iron formation, has been reported from eugeosynclinal (Minas Gerais, Postmasburg-Kalahari), miogeosynclinal (Maliy Khingan) and platform type (Morro do Urucum) tectonic set up. Non-volcanogenic manganese deposits of limestone-dolomite formation have likewise been reported both from eugeosynclinal (Usinsk and Apalachian deposits) and platform types (Ulu Telyaksk).

The volcanogenic-sedimentary manganese deposits can likewise be subdivided according to the association of rock formations in which they occur. In this subdivision the volcanic-siliceous facies of volcanogenic-sedimentary type of rock formations has only been considered as manganese deposits are reported only from this facies. The volcanogenic manganese desposits of volcanic-siliceous facies may be subdivided (on the basis of rock association), keeping SHATSKIY'S [1964] classification almost in its entirety. Only the "Remote Siliceous Formations" type is SHATSKIY'S classification has not earned the confidence of all workers and even according to SHATSKIY, its identity can only be established with difficulty. So this type should not be included as an unequivocal type in the classification.

Considering all aspects, manganese formations can be classified both on genetic and associational (with characteristic rock formations) basis. It has already been pointed out that the different associational types of manganese deposits cannot always be explained by any common genetic scheme. For example the limestone — dolomite formation contains manganese deposits of both volcanogenic (Morocco) and non-volcanogenic type (Usinsk, Ulu Telyaksk). Tectonic setting is also of little genetic consequence in many places. Though in eugeosynclinal types manganese deposits commonly show volcanic affiliation, unequivocal non-volcanogenic deposits are also contained in them. (cf. Usinsk deposit, U.S.S.R., Minas Gerais, Brazil.) Similarly the platform type deposits are generally non-volcanogenic (cf. Chiatura, Nikopol,. U.S.S.R) though evidences of volcanism and derivations of manganese ores thereform are also found (Marocco).

TABLE 3

Genetic-associational classification of manganese formations

MANGANESE FORMATIONS

ON GENETIC BASIS

- **HYDROTHERMAL**
  EX.: BUTTE, MONTANA, PIOCHE, NEVADA, LEADVILLE, COLORADO, SILVER CITY, NEW MEXICO (U. S. A). INAKURAI-SHI, HOKKAIDO (JAPAN). DZHEZDA AND NAYZATAS, CENTRAL KAZAKHSTAN (U. S. S. R).
- **SEDIMENTARY**
  - **NONVOLCANOGENIC**
    - ON THE BASIS OF ROCK ASSOCIATION
      - ASSOCIATION WITH ORTHOQUARTZITE-GLAUCONITE-CLAY & ORTHOQUARTZITE-CARBONATE FORMATION. MAINLY PLATFORM & MIOGEOSYNCLINAL TYPE. EX.: NIKOPOL, BOL'SHE TOKMAKSK, CHIATURA, LABINSK (U. S. S. R.). ORE DEPOSITS ASSOCIATED WITH GONDITE (INDIA).
        - OXIDE-CARBONATE ORES EX.: NIKOPOL, CHIATURA (U. S. S. R.).
        - OXIDE ORES EX.: MADHYA PRADESH, MAHARASHTRA, GANGPUR (INDIA).
      - ASSOCIATION WITH IRON FORMATION. EUGEOSYNCLINAL: EX.: MINAS GERAIS & BAHIA (BRAZIL); POSTMASBURG & KALAHARI (AFRICA). MIOGEOSYNCLINAL: EX.: MALIY KHINGAN (U. S. S. R.); PLATFORM TYPE: EX.: MORRO DO URUCUM, MATTO GROSSO (BRAZIL).
        - CARBONATE-SULFIDE ORES EX.: MINAS GERAIS (BRAZIL).
        - OXIDE-CARBONATE ORES EX.: MALIY KHINGAN (U. S. S. R.)
        - OXIDE ORES EX.: POSTMASBURG (S. AFRICA).
      - ASSOCIATION WITH LIMESTONE-DOLOMITE FORMATION. EX.: USINSK (U. S. S. R.).
        - CARBONATE ORES EX.: USINSK (U. S. S. R.)
  - **VOLCANOGENIC**
    - ON THE BASIS OF ROCK ASSOCIATION
      - GREENSTONE-SILICEOUS GROUP
        - GREENSTONE FORMATION (SPILITE, KERATOPHYRE, DIABASE ETC.). EX.: OLYMPIC PENINSULA (U. S. A.); S. URAL (U. S. S. R.); SREDNEGORSK (BULGARIA); WESTERN ALPINE & PENINE OPHIOLITIC ZONE (ITALY & SWITZERLAND).
        - JASPERIOD FORMATION EX.: S. URAL (U. S. S. R.); PARSETTEN & FALETTA DEPOSITS IN PENINE ZONE; GRAUBÜNDEN CANTON (SWITZERLAND); CHEVLYANOVICH (BOSNIA); FRANCISCAN FORMATION (U. S. A.).
        - SILICEOUS-SHALY FORMATION EX.: KELLERWALD & HARZ MOUNTAINS (GERMANY); HUELVA PROVINCE (SPAIN); FORTUNA HARBOUR (N. NEWFOUNDLAND); MOLUCCA PENINSULA (MALAYA) NSUTA DEPOSIT, BIRRIM SYSTEM, (GHANA).
      - PORPHYRY-SILICEOUS GROUP
        - PORPHYRITIC FORMATION (QUARTZ PORPHYRY, DACITE, ANDESITE AND LAVAS OF INTERMEDIATE COMPOSITION). EX.: GLIB-EN-NAM (MOROCCO); KOLNINGSBERG, LÅNGBAN (SWEDEN).
        - SILICEOUS-SHALY FORMATION. EX.: CENTRAL KAZAKHSTAN (U. S. S. R.).
- **SUPERFICIAL**
  EX.: SUPERGENE OXIDE ORES OF GHANA, GABON (AFRICA); MINAS GERAIS, (BRAZIL); DONGRI BUZURG (INDIA) ETC.

It is, therefore, necessary to evolve a genetic-associational classification of manganese formations that may include the most important deposits in its folds. It is understood, however, that no individual associational type is a sole representative of an unique genetic class. More than one associational type may characterize a genetic class and some particular associations may be product of any of the different genetic classes.

The classification of manganese formations suggested by the present author is given in Table 3.

CONCLUSION

An attempt has been made in the preceding pages to synthesize the data on the mode of genesis of the principal maganese deposits of the world an their association with characteristic rock formations. The accumulated data under review indicate that a broad-based three-fold genetic classification (hydrothermal, sedimentary and superficial) encompasses most of the important manganese deposits of the world. Of these three genetic types, the sedimentary deposits are, by far, the most important and can, genetically, be further subdivided into non-volcanogenic and volcanogenic types. The non-volcanogenic and volcanogenic-sedimentary deposits are associated with characteristic rock formations. In the case of the latter, clearcut divisions may be made into Greenstone — Siliceous and Porphyry— Siliceous Groups according to the association of characteristic volcanic rocks. The Greenstone — Siliceous Group can further be subdivided into *(i)* Greenstone formation, *(ii)* Jasperoid formation, and *(iii)* Siliceous — Shaly formation and the Prophyry — Siliceous Group into *(i)* Porphyry formation, and *(ii)* Siliceous — Shaly formations. All these formations merge into one another and are genetically related. The non-volcanogenic — sedimentary formations can be subdivided, on the basis of rock association, into *(i)* association with orthoquartzite — glauconite — clay and orthoquartzite — carbonate formations, *(ii)* association with iron formation, and *(iii)* association with limestone — dolomite formation.

The various subdivisions on the basis of rock association cannot always be related by genesis and/or tectonic set up. Thus, the volcanogenic deposits are generally found in geosynclinal and the non-volcanogenic deposits in platform type basins, though there are evidences on the contrary e.g. the platform type deposit at Morocco is volcanogenic whereas the eugeosynclinal deposits at Minas Gerais and Usinsk are non-volcanogenic. Similarly rock associations do not unequivocally characterize a volcanogenic or non-volcanogenic manganese deposit e.g. manganese deposits with limestone dolomite association at Usinsk and Ulu Telyaksk (U.S.S.R.) are non-volcanogenic whereas those at Morocco are volcanogenic. Thus no individual associational type is a sole representative of an unique genetic class. More than one associational type may characterize a genetic class and some particular associations may be product of any of the different genetic classes.

Finally, the sedimentary manganese deposits of different genetic and associational types are characterized by oxide and carbonate and rarely sulphide ores, sometimes exhibiting lateral variation. Such mineralogical — chemical manifestations of the ores may either reflect differences in depositional environments (controlled by Eh and pH) or the post-depositional diagenetic changes.

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THE POLLEN GRAINS OF THE CARBONATE MANGANESE ORE OF THE SHAFT III. IN ÚRKÚT

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INTRODUCTION

We have previously published in papers the plant microfossils found in oxide manganese ore [SIMONCSICS & KEDVES 1961] and the *Pteridophyta* spores of carbonate manganese ore in Úrkút [KEDVES & SIMONCSICS 1964 *a*, *b*]. In our works we have put forward the idea that the oxide and carbonate ore could be formed sooner than the black coal of Liassic in Mecsek and even sooner than the age of Upper Liassic. On the basis of the main spore- and pollen-types being present in the carbonate manganese ores we have made attempts to reconstruct the vegetation of the sedimental recess and the surrounding area and according to the microfossils we have divided the carbonate ore deposit into level A *(Classopollis)*, level B *(Crassosphaeridae)* and level C *(Spheripollenites)*.

In the present paper we are going to demonstrate the pollen-types of the carbonate manganese ore deposit in Úrkút together with their botanical and stratigraphical consequences.

SYSTEMATIC PALYNOLOGY

Although the descent of the Jurassic disperse pollen-types is uncertain, we are not going to set up an artificial pollen-system, but we are trying to enlist the pollen-types on the basis of a natural relationship instead. In our systematization we are following Soó's work [1963] to an extent where the degree of relationship is certain, probable or possible.

Phylum: **Gymnospermatophyta**
Subphylum: **Pteridospermophytina**
Classis: **Pteridospermopsida**
Ordo: **Caytoniales**

Vitreisporites pallidus [REISSINGER, 1938] NILSSON 1958 (Plate II. *27*, *28*).

The pollen-type is rare in the manganese ore of Úrkút and can be found only in the level A of the carbonate deposit. Its stratigraphy is given from Jurassic to Lower Cretaceous by COUPER [1958]. There is no recent equivalent to this pollen-form. From the fossil disperse pollen-types the type *V. signatus* LESCHIK 1955 from Keuper, the type *V. bjuvensis* NILSSON 1958 from Rhaetic, the type *Caytonipollenites contectus* DE JERSEY 1959 and *C. subtilis* DE JERSEY 1959 from Lower Jurassic (?) and the type

Caytoniales-Pollenites diaphanus PAUTSCH 1958 and *Caytoniales? Pollenites fuscocorpus* PAUTSCH 1958 from Keuper show similarity to the type of Úrkút.

The similarity of the *Vitreisporites pallidus* to the *Caytonianthus arberi* [THOMAS] HARRIS and *C. oncodes* HARRIS pollens of associated type was shown by COUPER [1958]. Its similarity to the *Sagenopteris nilssoniana* [BRONGN.] WARD pollen is also worth mentioning [HARRIS 1926, 1932].

According to the above-mentioned facts the descent of this pollen-form is given in the *Caytoniales.*

It is revealed in Soó's system [1963] that the *Caytoniales* — beside the *Pteridospermales* — is an order of the *Pteridospermophytina* and of the class *Pteridospermopsida.* According to ENGLER'S Syllabus [1954] the *Caytoniales* which is the second order to the *Cycadopsida (Cycadophyta)* — descending from the *Pteridospermae* — occured from Upper Triassic to Upper Jurassic.

Classis: **Cycadopsida**
Ordo: **Cycadales**

Cf. *Cycadopites* [WODEHOUSE, 1933] ex WILSON & WEBSTER 1946 fsp. (Plate II. *21, 22*).

The plicated pollen-form found in the only sample does not make possible a diagnosis. The sample possibly belongs to the *Cycadopites* fgen.

Description: It is a monocolpate pollen. Its contour is ellipsoid, but one of the two poles is somewhat blunter. Its colpus extends from pole to pole and near the poles it widens out. The wall is thick, double-layered, about 5μ. The ectexine-endexine rate is 4:1. The exine is smooth-chagrenate, here and there corroded. Its length is 60μ, its width is 42μ. The pollen-grain is plicated, and because of the fold running along the colpus there is some similarity to the genera *Eucommiidites* ERDTMAN 1948 and *Bennettitaepollenites* [THIERGART, 1949] R. POT. 1958.

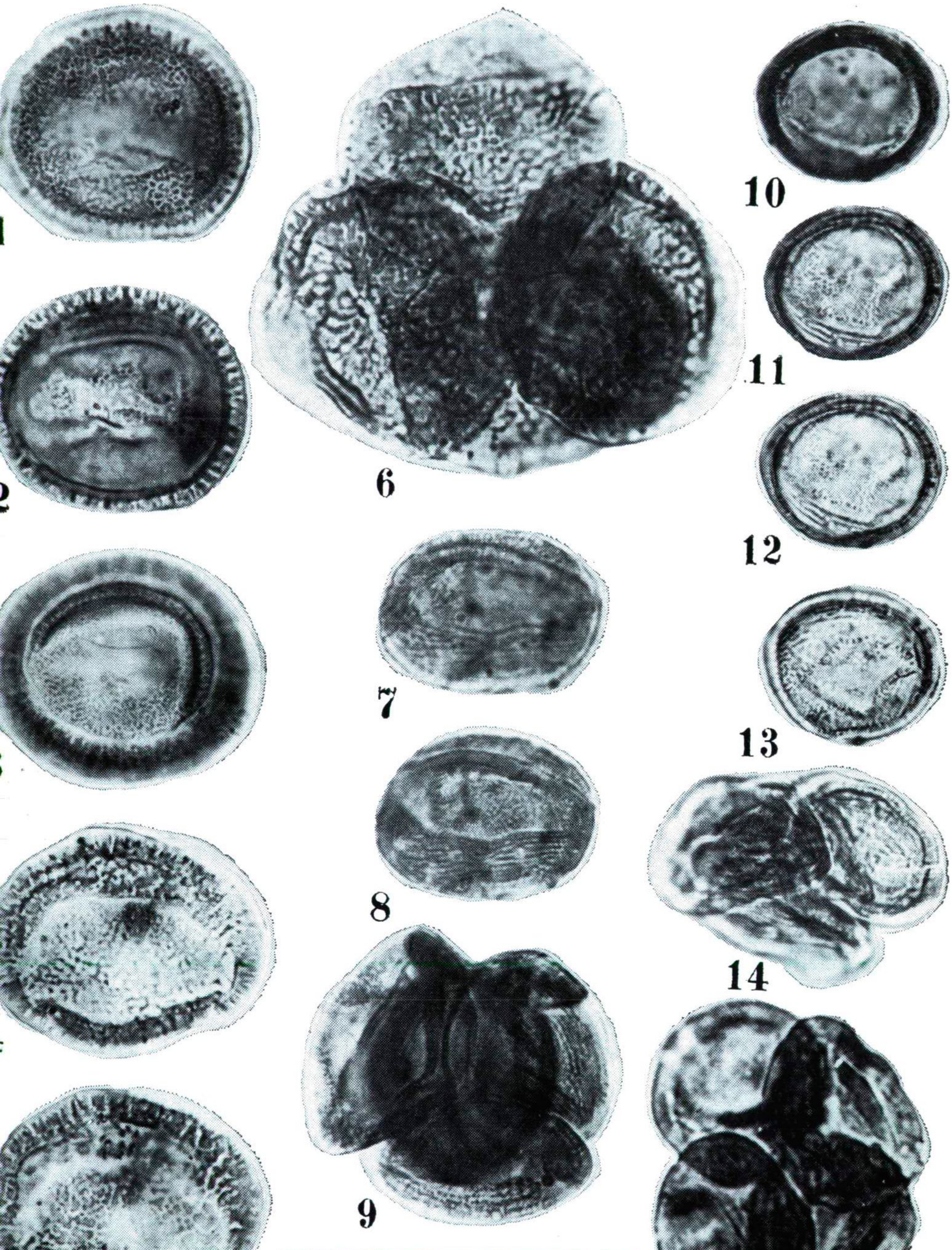
The holotype of the *Cycadopites* [*C. follicularis* WILS. & WEBST., 1946] descends from Tertiary deposits. Not neglecting the fact that the *Cycadales* occurs from Traissic on and its maximum of development can be dated in Jurassic and Lower Cretaceous, the descent of this pollentype from *Cycadales* is probable. The *Pentoxylales* — classed as *Cycadopsida* by Soó [1963] — cannot be considered as a pollen-producing taxon, because it was found only in India, whereas, on the other hand, the *Nilssoniales* — classed as belonging to the same order — can be regarded as a pollen-producing taxon, because it was at its height in Jurassic. But we do not know the pollen-type of *Nilssoniales,* because it has died out and we have no knowledge of its associated pollen, either.

Classis: **Cycadopsida or Ginkgopsida**

Monosulcites minimus COOKSON 1947 ex COUPER 1958 (Plate II. 9—11).

The pollen-type consistently occurs everywhere in the carbonate deposit and in the basis of the deposit (the lower part of the level A) it is the dominant form of the spore-pollen complex.

1—3. — *Classopollis grandis* n. fsp.
4—6. — *Classopollis grandis* n. fsp.
7—9. — *Classopollis classoides* [PFLUG, 1953] POCOCK & JANSONIUS 1961.
10—14. — *Classopollis minor* POCOCK & JANSONIUS 1961.
15. — *Classopollis classoides* [PFLUG, 1953] POCOCK & JANSONIUS 1961.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The origin of the exines is uncertain. From the recent classes — beside the *Ginkgopsida* — the *Cycadopsida* produces similar pollen-types. Concerning this problem literary data were published previously [SIMONCSICS & KEDVES, 1961].

From the associated spores COUPER [1968] records the *Andostrobus manis* HARRIS and *A. wonnacotti* HARRIS prepared pollen-grains which correspond to the formerly examined *M. minimus* COOKSON. On the other hand the above-mentioned two associated *Androstrobus* pollens are compared with the Permian disperse *Gynkgaletes* LUBER 1955 fgen. by POTONIÉ [1958]. Regarding the habit the *Androstrobus* corresponds to the *Cycadales*. According to THOMAS & HARRIS [in GOTHAN & WEYLAND, 1964] the male flowers of the *A. manis*, the *Nilssonia compacta* and the *Beania gracilis* on one hand, and on the other the *A. wonnacotti*, the *N. tenuinervis* and the *B. mamayi* belong together and are to be classed as *Nilssoniales*. The *Nilssoniales* is an order of the *Cycadopsida*.

Classis: Cf. **Cycadopsida**

Monosulcites urkutiensis SICS. & KDS. 1961 em.

The species was recorded by the authors from the oxide manganese ore deposit. Instead of the holotype of the species only the holotypes of two subfsp. found in the manganese ore deposit were given and the naming of the first subfsp. was not correct, either, so the species is to be emended.

Diagnosis: see SIMONCSICS & KEDVES [1961].
Holotype: SIMONCSICS & KEDVES [1961] Pl. II. *24, 25.*
Locus typicus: Úrkút.
Stratum typicum: oxide manganese ore, Middle Jurassic.

Note: The relationship of the pollen-grain can be named as *Cycadopsida*. Though monocolpate character occurs both in the *Ginkgopsida* and in the *Monocotyledonopsida*, it cannot be *Monocotyledonopsida* because of its age and cannot be *Ginkgopsida*, either, first of all because of its shape. The pollen of the *Ginkgopsida* is more spindle-shaped very often with sharp apices. Whereas one or both apices of the *M. urkutiensis* are blunted, sometimes even circular.

The subspecies together with the form found in the carbonate manganese ore deposit are as follows:

a) subfsp. *urkutiensis* nov. nom.
Syn.: subfsp. *hyalinoides* SICS. & KDS. 1961.
This subfsp. has not occured in our present material.

b) subfsp. *scabratus* SICS. & KDS. 1961 (Pl. II. *8, 7*). There were only a few samples in the basis of the deposit.

c) subfsp. *circulus* n. subfsp. (Pl. II, *1, 2, 5, 6*).
Diagnosis: The contour is approximately circular. The colpus is short, not longer than the 2/3 part of the pollen-diameter. There are plications of exine on both sides of the colpus which are bifurcating at one end of the colpus and are not or, at least, in a lesser degree are at the other.

Maximal size: 16 μ (15—22 μ)
Holotype: Pl. II. *5, 6;* prep. U—III—28—95—3.
Locus typicus: Úrkút.
Stratum typicum: carbonate manganese ore, Middle Jurassic.

Note: This subfsp. can be differentiated from the *M. urkutiensis scabratus* by its shape, from the *M. urkutiensis urkutiensis* by its ornamentation. References to its relationship, see above. For the sake of exactitude on the basis of the colpus and the plications we have to mention the similarity with the associated spores of the genus *Androstrobus* and, on the other hand, we have to consider its relationship to the *Ginkgopsida* to be improbable for the reason of its shape.

Classis: **Chlamydospermophytina**
Ordo: **Bennettitales**

Bennettitaceaeacuminella cf. *simplex* MALYAVKINA 1953. (Pl. II. *23—25*).

The only sample found in our material — according to descriptions, drawings and reproductions available — on the whole corresponds to the species recorded from Rhaetic sediments of Saghiz with the possible difference that our sample is not intragranulate, but granulate. POTONIÉ [1958] refers to the fact in Synopsis II, that the *B. simplex* is similar to the associated pollen of the *Wielandiella punctata* NATHORS. This pollen-type offers newer data to the abundant *Bennettitales* flora in our area.

Beside the Rhaetic period this type occured in former Liassic coals, too, recorded by ROGALSKA [1954]. So the stratigraphy of this form can be given in the Rhaetic-Liassic—Middle Jurassic.

Classis: ? **Chlamydospermophytina**
Ordo: ? **Bennettitales**

Eucommiidites ERDTMAN 1948.

On the basis of COUPER'S profound analysis [1958] and the corresponding parts of POTONIÉ'S Synopsis II the genus is not typically tricolpate, but praecolpate. If it is so, the former hypothesis that the producers of the abovementioned pollen-genus were angiospermal, must be dropped. This statement is in accordance with many recent opinions.

There is neither equivalent, nor — at least — similar pollen among the recent *Gymnospermae*. But we find pollengrains with similar structure among the associated pollentypes of the extinxt *Gymnospermae*, namely in the *Pteridospermales* and the *Bennettitales*. We are forced to drop the *Pteriospermales* as a producing taxon, because — according to our recent knowledge — it died out already in Triassic and its microspores were of enormous size (200—500 μ). The most probable order is the *Bennettitales*. The probability is increased by the fact that its heyday was in Jurassic, when it flourished in great number according to WIELAND [1916].

The known pollen-types of the *Bennettitales* — with the acceptance of the trilete microspores of the genus *Cycadocephalus* NATHORST — are monocolpate with a touch of praecolpate. It means that two or more plications with or without furrow run parallel with the main colpus. In case of two additional furrows the pollen seems to be tricolpate and types similar to *Eucommiidites* arise.

Eucommiidites troedssonii ERDTMAN 1948 ex COUPER 1958. (Plate II. *12—14, 15, 16, 26*)

This pollen-form can be found in all three levels with a few samples. In one case (Plate II. *26*) a major heap of pollens occurs in the preparation. The praecolpate character can be well seen even in the case of these presumably immature grains.

Eucommiidites rugulatus n. fsp.
(Plate II. *17, 18, 19, 20*)

Diagnosis: The contour of equator is oval with somewhat angular poles. In the longitudinalis axis of the aquator a thin colpus runs. The rate of the longer diameter equatorial and the colpus is 4:3. The colpus is not opened at the ends. A 2 μ wide plication goes along with the colpus in the sample of the type. In the opposite side two plications — growing narrow at the poles while they reach the equator — run parallel with the equator and its longitudinal axis. The wall is two-layered tectate-rugulate-baculate. It is thin (under 1 μ). Maximal size: 23×14—28×18 μ.

Holotype: Plate II. *17*, *18*; prep. U—III—8—116.
Locus typicus: Úrkút.
Stratum typicum: carbonate manganese ore, Middle Jurassic.

Note: In its structure this pollenform is similar to the species of *Schopfipollenites* POTONIÉ & KREMP 1954, but without "umbo". From the other species of the *Eucommiidites* form-genus it can be distinguished by its closed colpus, shape and ornamentation of the wall. In level A it occured only in a few samples together with the *E. troedssonni.*

Classis: ? **Coniferopsida**
Ordo: ? **Taxales**

Spheripollenites subgranulatus COUPER 1958 (Plate II. *3*, *4*)

Beside the *Classopollis* the pollen-type *Spheripollenites* occurs en masse, the spectra are dominated by them especially in the upper level C. The probability of its connection with the *Taxales* — beside morphological similarities — is backed by the fact that the macrofossils of the order appear in greater number from the Jurassic period. There are similar samples to the *Spheripollenites* among the *Taxodiaceae* pollen-types, too, but positive representatives of the family are hardly known before Cretaceous (exception: a *Sciadopitys* leaf from Lower Cretaceous. The *Cupressaceae* can also be regarded as a pollen-producer to the family, but its microfossils are known only from Upper Jurassic.

The stratigraphy of the pollen-form is given by COUPER [1958] from Middle Jurassic. to Upper Cretaceous.

Gymnospermatophyta incertae sedis

Classopollis [PFLUG, 1953] POCOCK & JANS. 1961.

After some imperfect description of the morphology of the pollen-genus [PFLUG, 1953, BALME, 1957, COUPER, 1958, ZAUER & MTCHEDLISHVILI, 1954 and others] this problem was cleared by POCOCK & JANSONIUS 1961. They named the *C. classoides* [PFLUG, 1953] POCOCK & JANSONIUS 1961 as genus type. COUPER [1958] emended the

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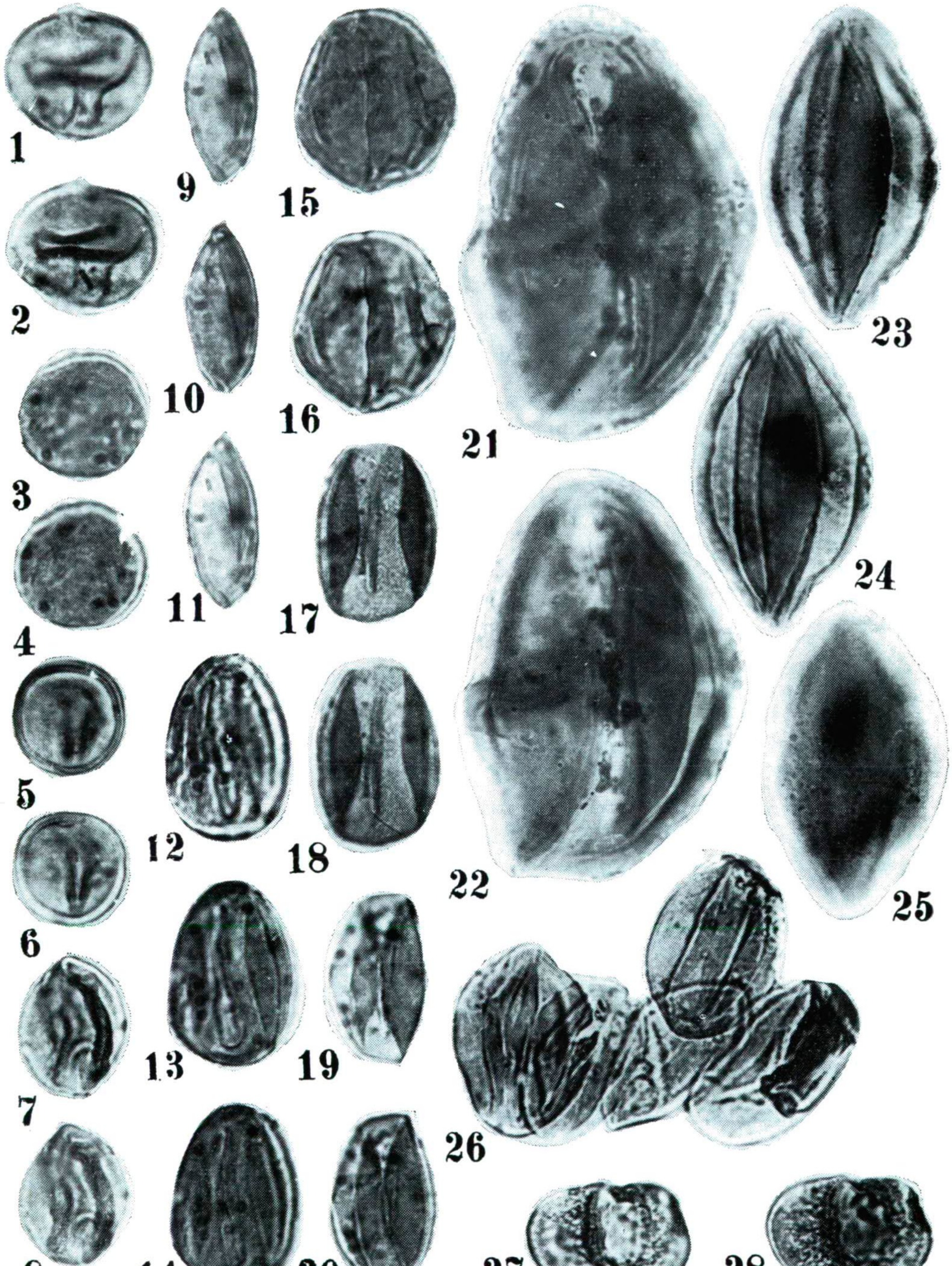
1—4. — *Spheripollenites subgranulatus* COUPER 1958.
5,6. — *Monosulcites urkutiensis* subfsp. *scabratus* SICS. & KDS. 1961.
7,8. — *Monosulcites urkutiensis* subfsp. *scabratus* SICS. & KDS. 1961.
9—11. — *Monosulcites minimus* COOKSON 1947 ex COUPER 1958.
12—16. — *Eucommiidites troedssonii* ERDTMAN 1948 ex COUPER 1958.
17—20. — *Eucommiidites rugulatus* n. fsp.
21,22 — Cf. *Cycadopites* fsp.
23—25. — *Bennettitaceaeacuminella* cf. *simplex* MALYAVKINA 1953.
26. — *Eucommiidites troedssonii* ERDTMAN 1948 ex COUPER 1958.
27,28. — *Vitreisporites pallidus* [REISSINGER, 1938] NILSSON 1958.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former *Pollenites torosus* REISSINGER 1950 by including every disperse *Classopollis* in the species *C. torosus.* POCOCK & JANSONIUS [1961] pointed out that the inclusion of several species in the *Classopollis* genus in highly reasonable, later BURGER [1965] emended the *C. torosus* [REISSINGER, 1950] COUPER fsp. beside other species belonging to the *Classopollis* and separated it from the *Classopollis classoides.* The present material contains several *Classopollis* form-species, which are partly described and partly newly discovered form-species.

The equivalent recent pollen-type of the *Classopollis* form-genus is not known. But on the basis of similarity of morphology and of certain elements of wall structure it might be in connection with *Gymnospermatophyta.* Identifications and similarities with pollen-grains clinged to cones or axis of cones and with grains not prepared from the pollen sac are not at all convincing. The connection of the *Classopollis* fgen. with the fossile genera *Cheirolepis, Brachyphyllum* and the *Pagiophyllum* is probable, but not certain.

The genus *Cheirolepis* SCHIMPER is *Coniferales* incertae sedis according to SEWARD [1964]. The genus *Brachyphyllum* BRONGN. can be found among the *Cupressaceae* at SEWARD [1964], in GOTHAN & WEYLAND'S book [1964] it is regarded as a *Coniferae* with questionable relationship and at ZIMMERMANN [1959] on account of the anatomy of epidermis it is placed into the *Araucuriaceae.* By the same author the genus *Pagiophyllum* HEER belongs to the *Araucariinae,* it is in relationship partly with the *Araucaria* and partly with the *Araucariaceae.* It is worth mentioning that the disperse pollen-grains of the *Classopollis* type are classed as genus *Brachyphyllum* or *Pagiophyllum* by several Soviet research-workers.

The genus *Classopollis* occurs in every sample of the carbonate manganese ore of Úrkút. It is the dominant pollentype of level A, greatly surpassed in level B, while it is going to be a common type again in level C, though its importance is subordinate to the *Spheripollenites* (and the *Monosulcites*). In the material of Úrkút several *Classopollis* form-species occur.

Classopollis classoides [PFLUG, 1953] POCOCK & JANSONIUS 1961 (Plate I. *7, 8, 9*)

There are numberless unharmed samples in our material. A part from the form-species *Classopollis torosus* [REISSINGER, 1950] COUPER 1958 published from the oxide manganese ore belongs to the species *Classopollis minor* POCOCK & JANSONIUS 1961 and the *Classopollis classoides.*

Classopollis minor POCOCK & JANSONIUS 1961
(Plate I. *10—13, 14*)

The type *C. minor* is not common in the carbonate deposit, but types under the name of *C. torosus* [REISSINGER, 1950] COUPER 1958 (published in our former paper on oxide ore, see SIMONCSICS & KEDVES [1961] Plate II. *1—5, 6, 11, 12, 21—22, 23, 24, 25*) can be mostly included in the above-mentioned species. The pollen-type — according to its authors — occurs from Lower Jurassic to Eocene.

Classopollis grandis n. fsp.
(Plate I. *1—3, 4, 5, 6*)

Diagnosis: The shape of the pollen-grain is spherical with a thick equatorial ring. The proximal hemisphere is gently conical, the distal one is more convex. The exine of hemisphere is about 1 μ thick. The distal exine is tectate, finely granulate, of rugulate structure and its contour is not clear, with a pore of 5 μ diameter. The distal pole is separated from the equatorial thickening by a tenuous exine layer, the

PFLUG-"rimula". In the proximal side a triangular thin exinal field of unclear contour refers to tetradic juncture. The proximal hemisphere is also tectate with coarsely circular or irregular ornamentations, which fuse into equatorially lengthened elements toward the equator and take part of formation of the ring. The equatorial exine thickening is 3—5 μ. The thick wall is dissected by channels running roughly parallel with the equator. The channels by flanking sections of thick, circularly or equatorially lengthened wall make an impression of endostriae.

The diameter of the typical sample is 50×47 μ, the size of the other samples is in the range 36—50 μ, the length of the polediameters is in the range 32—40 μ.

Holotype: Plate I. *1—3;* prep. U—III—3—82.

Stratum typicum: green, grey, finely streaked carbonate manganese ore, Middle Jurassic.

Locus typicus: Úrkút.

Note: The new species differs from all the other *Classopollis* not only in size, but also in the structure of the equatorial thickening, i.e. the striate character of it is not so distinct as in other *Classopollis*. This pollen-type occurs rarely, mostly in the upper level of the deposit.

SUMMARY

1. On the basis of the above systematic palynological data we have pointed out 4 or 6 classes of the phylum *Gymnospermatophyta* in carbonate manganese ore of Úrkút. The orders *Caytoniales*, *Cycadales* and *Bennettitales* may be considered as undoubtedly existing ones.

2. The relationship between the two dominant pollenspecies is uncertain, but the descent of the former from the *Araucariaceae* and the descent of the latter from the *Taxales* is possible.

3. On the basis of macrofossils the flora of Middle Jurassic is characterized by the widespread range of the *Bennettites*. This is not reflected in the pollen-spectra of the samples even in case, if we suppose that the genera of the pollen-species *Eucommiidites* come from the *Bennettites*. The possible insect-pollination could cause the low number of the pollens, the one-sidedness of the oecological conditions could give an explanation to the poverty of genera.

4. The systematic analysis of the pollen-material did not change our former stratigraphical conclusions. The age of the carbonate deposit is to be dated in Middle Jurassic.

5. The picture of the Middle Jurassic flora surrounding the recess accumulated with manganese in Úrkút has remained unchanged and corresponds to the results published in our former paper [KEDVES & SIMONCSICS, 1964].

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METASOMATIC DOLOMITIZATION ON THE WESTERN PART OF THE NAGYSZÁL MOUNTAIN

GY. VITÁLIS and J. HEGYI-PAKÓ

Concerning the exploration for binding materials made in the years 1965—1967, the western part of the Nagyszál was explored by several drill-holes. During these exploration works several hydrothermal traces pointing to an ancient thermal spring activity were observed both in the present limestone quarry and the materials of drillings. Significant dolomitization of the limestone series presumably connected with hydrothermal metasomatism could be established. The thermal spring acitivity, almost in the whole area — mostly on the 420 and 450 m levels — is marked partly by numerous calcite veins partly by typical spring-caved spherical chambers and galleries opened with the mining. The calcite veins refer to a former spring activity, the spring-cavern galleries to the last uprush places of the one-time springs. This uprush might be contemporaneous with the end of the spring activity.

In the Upper Liassic limestone series, opened in 200 m thickness in some places, limonite, pyrite and kaoline also of hydrothermal origin were found beside several variously coloured calcite veins. The impregnation of fluidal character and sugar-like appearance of the rock texture were observed in several places of the limestone series. These rocks represent in most cases transitional rock types between the limestone and dolomite, as it is shown by laboratory investigations.

As known from the literature, in the underlying strata of the Noric limestone series of the Nagyszál, reaching more than 200 meters, Karnic dolomite is deposited in a thickness of several hundred meters. Smaller spots of this dolomite crops out in the neighbourhood of the borehole XV-1 *(Fig. 1)*. The Karnic dolomite is considered as chemical sediment of marine origin.

The appearance of dolomitic limestone, limey dolomite and dolomite observed in the boreholes crossing the Noric limestone strata, points to hydrothermal veins. On the basis of the hydrothermal traces observed in the area as well as the vein-like or stocky appearance of dolomitic rocks, they are considered of metasomatic origin. Of the dolomitized strata traversed by boreholes, the average thickness of dolomitic limestone is 3,45 m, that of the limey dolomite 2,77 m and finally the average thickness of the dolomite is 1,0 m.

The one-time hydrothermal activity may be considered as a process connected with the andesite volcanism of Visegrád, Nagybörzsöny and the Miocene volcanism of the Cserhát. The hydrothermae rushed up mainly along transverse fractures of N—S and NW—SE direction. The hydrothermae percolating through the carbonate rocks resulted in their dissolution of different rate and under favourable tectonical conditions, where the percolating water could move more freely, dolomite could formed from the solution. The dolomitization is a fairly complicated process, whose rate depends mostly upon the temperature, CO_2 pressure, the pH and the concentration of ions.

The dolomitization connected with hydrothermal metasomatism and the dolomitic impregnation of the limestone strata, respectively, is represented by manysided investigations carried out on the rock materials of the region.

The primary, Karnic dolomite of marine origin, shown in *Fig. 2,* is of equigranular texture, with grain sizes of 100—200 μ. The typical Noric limestone *(Fig. 3),* is mainly of fine-crystalline texture. In *Fig. 4* a dolomite veinlet intruded into the limestone is to be seen. The metasomatic limey dolomite shown in *Fig. 5* is of inequigranular texture, with preponderance of the fine-grained fraction (70%). In the fine-grained matrix middle- (20%) and coarse-grained (10%) calcite is to be seen.

To illustrate the surface of the primary, sedimentary and that of the metasomatic dolomite, also electron microscopic investigations were made. *Fig. 6* shows the electron photomicrograph of the surface of the primary, marine and *Fig. 7* that of the secondary, metasomatic dolomite. The electron microscopic investigations were made by carbon replica of fresh fracture-surface of the rock.

Comparing the photomicrographs, the most remarkable is the appearance of parallel edges on the surface of marine dolomite, oriented in one direction within the phase boundaries. The surface is generally of "foamy" character. The lines of metasomatic dolomite are undetermined, the surface is relatively smooth. On both photomicrographs dolomite rombohedrons remained as dust on the fracture-surface are to be seen.

The map of hydrothermally decomposed and dolomitized rock types in the Nagyszál area explored by drillings is shown in *Fig. 1.*

The map shows the greater — 0,8—3,7 m in thickness — calcite veins and thermal spring-caved galleries according to the conditions in the year 1968. The percentage of the hydrothermally decomposed and the dolomitic rock types compared to the thickness of the drilled strata is shown by the circle diagrams drawn in the place of the exploration boreholes. The comparison between the surface and 420 m level (the lowest mining level of the "Danube Cement and Lime Work's" quarry) is illustrated by the inner circle of the diagrams whereas the situation between 420—380 m levels is represented by the outer circles. The explanation of *Fig. 1* is at the same time a summary wherein the data of boring VII-04 traversing the lower strata (between 352,7—272,7 m) denoted by hexagon are not shown.

Evaluating the data of *Fig. 1* from the point of view of metasomatic dolomitiza-

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Fig. 2. Photomicrograph of primary dolomite from the depth of 90,0 m, hole XV-1. Crossed nicols, $\times$50.

Fig. 3. Photomicrograph of limestone from depth of 50,0 m, hole VIII-3. Crossed nicols, $\times$*50.*

Fig. 4. Photomicrograph of dolomite veined limestone, from the depth of 50,0 m, hole VIII-3. Crossed nicols, $\times$50.

Fig. 5. Photomicrograph of secondary limey dolomite from the depth of 130,0 m, hole VIII-3. Crossed nicols, $\times$50.

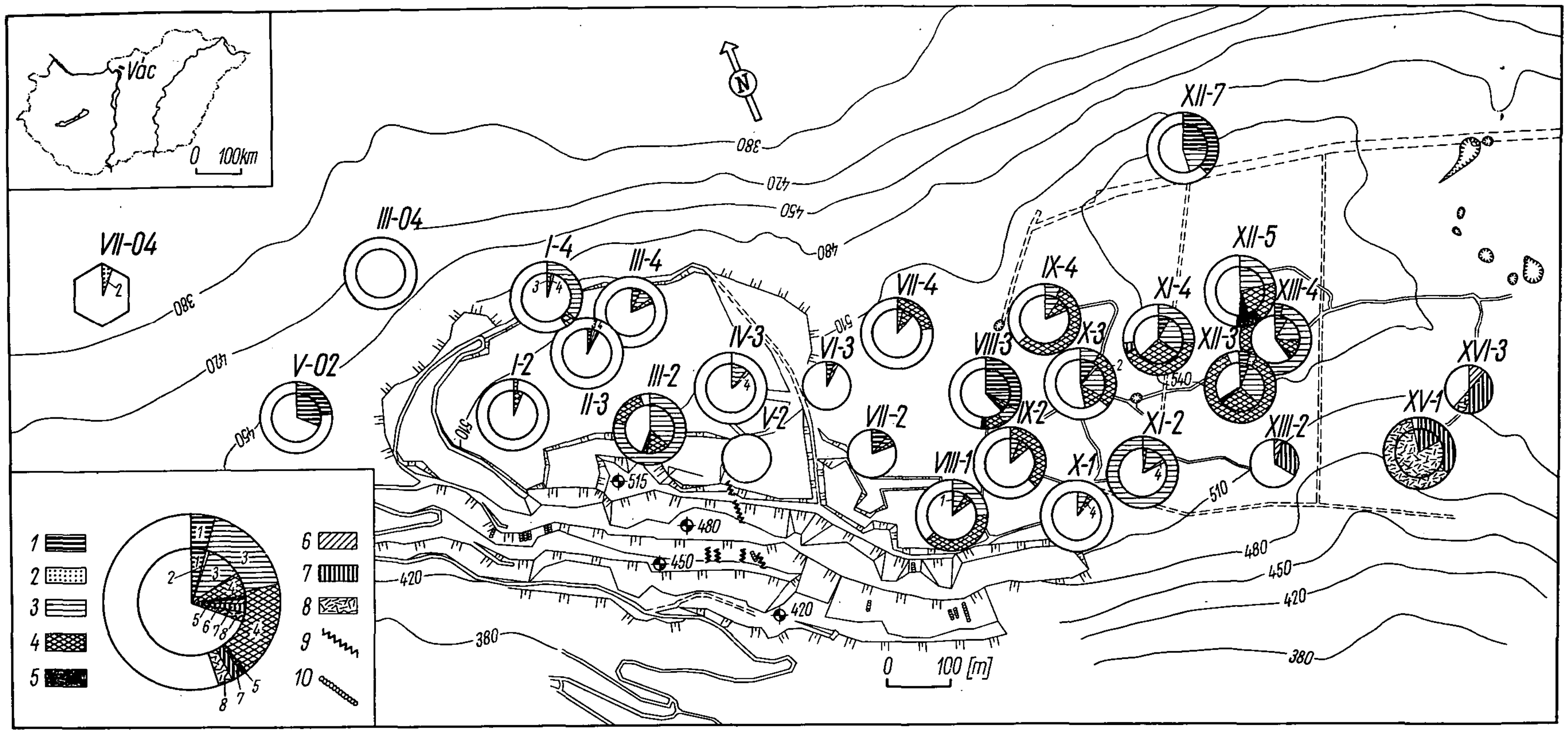
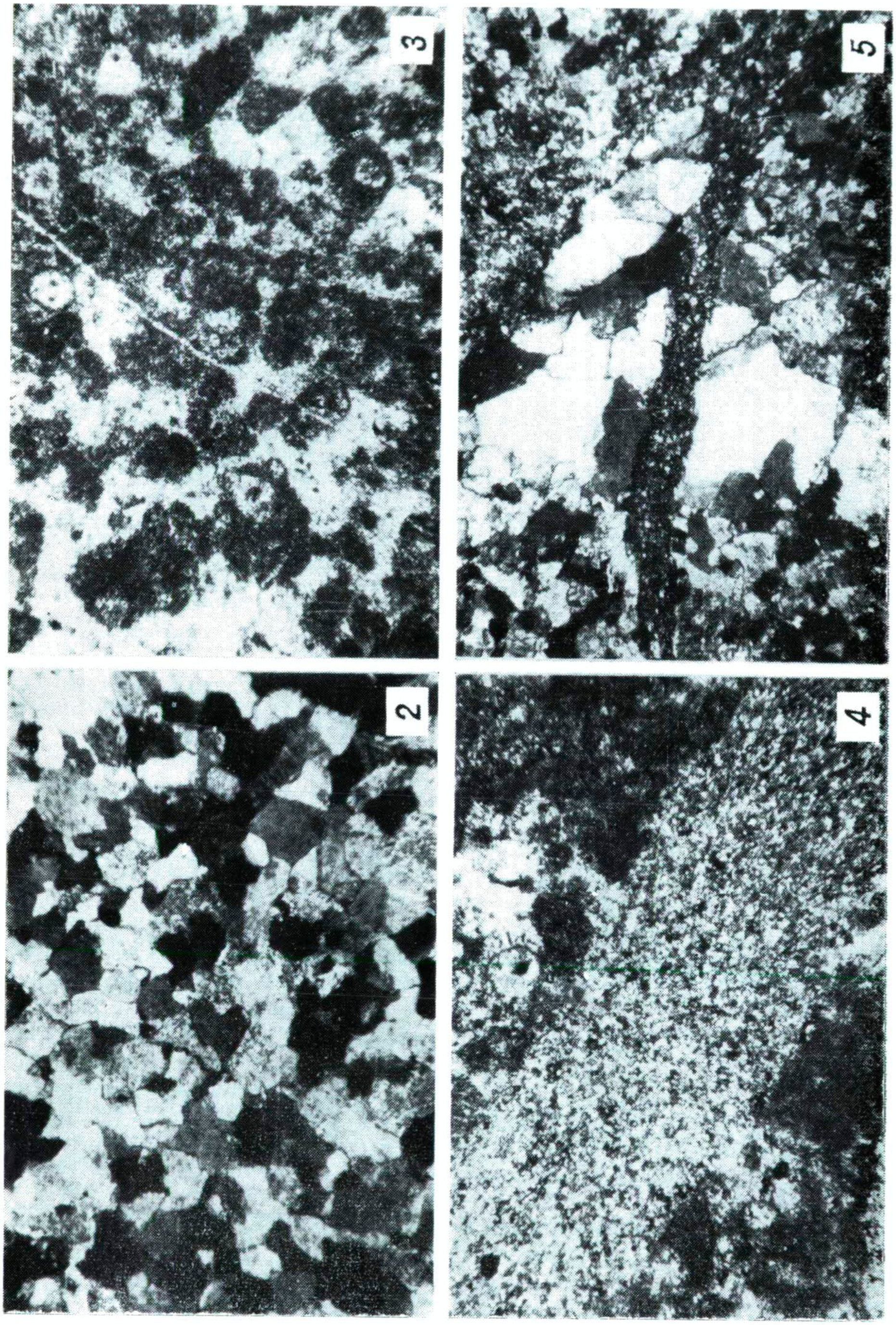


Fig. 1. Map of the hydrothermally decomposed and dolomitized rock type tse hon western part of the Nagyszál at Vác

Explanation and summarizing diagram (in the left lower corner of the figure): The inner circle represents the opened up part from the surface to the level 420 m; the outer circle the levels betwen 420—380 meters; the hexagon indicates the levels between 352,7—272,7 meters. 1) Hydrothermally decomposed limestone; 2) Hydrothermally decomposed silty rock flour, silt with rock flour; 3) Dolomitic limestone; 4) Limey dolomite; 5) Dolomite (secondary); 6) Dolomitic limestone; 7) Limey dolomite; 8) Dolomite (primary); 9) Larger calcite vein; 10) Cave system of thermal spring origin.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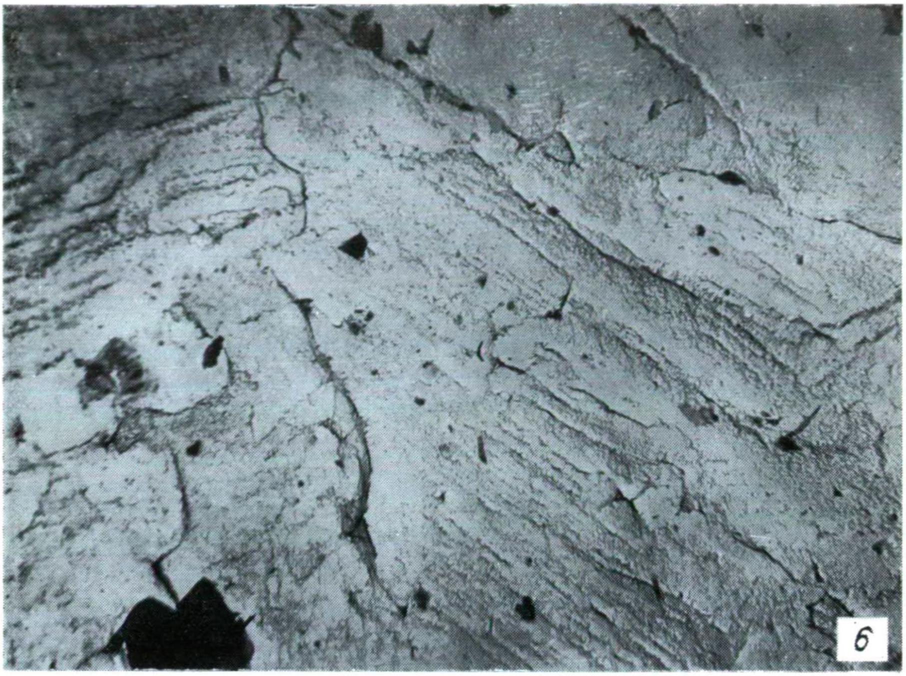

Fig. 6. Electron photomicrograph of the surface of primary dolomite, from the depth of 90,0 m, hole XV-1. ×4800.

tion the followings are emphasized. The percentage of the secondary dolomite types compared to the total thickness of strata crossed by boreholes is as follows:

	Between the surface and 420 m level	Between the levels 420—380 m
Dolomitic limestone	11,17%	17,41%
Limey dolomite	7,24	17,16
Dolomite	0,42	0,65

As it is shown by these values the metasomatic dolomitization resulted in mainly dolomitic limestone from the surface to the depth of 420 m level and dolomitic limestone and limey dolomite in equal parts between levels 420—380 m, respectively.

Summarizing the results is to be seen that the area NW from the line of boreholes VIII-1—VIII-3 is less dolomitized. A more considerable dolomitization is found only in the strata of borehole III-2, consisting of mainly dolomitic limestone. This part of the area belongs namely to the line of "active" fractures which is shown also by the nearness of a great calcite vein opened up in the quarry, referring to later hydrothermal actions.

The rate of dolomitization is more expressed reagionally from W to E and vertically toward the greater depths. The uprush places of hydrothermae resulting in dolomitization may be supposed where the secondary dolomit types are the thickest. These places agree well with the greatest fractures shown on the geologic map of this region.

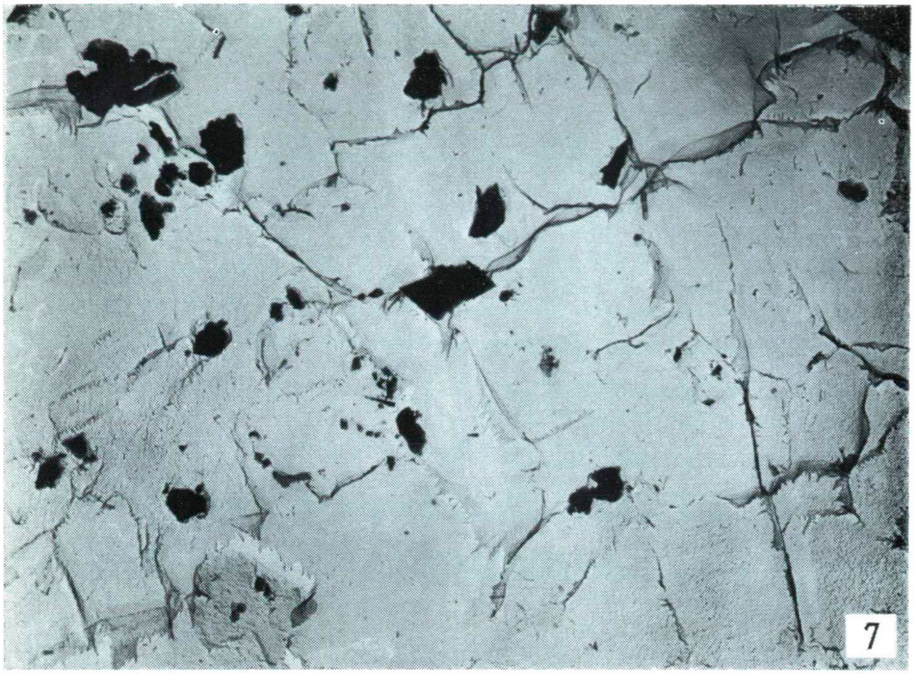

Fig. 7. Electron photomicrograph of the surface of secondary limey dolomite from the depth of 130,0 m, hole VIII-3. ×4800.

The stronger dolomitization of the eastern part may be connected with the transverse fracture between the boreholes VIII-1 and VIII-3 as well as IX-2 and IX-4, further with the fracture SE from the line of boreholes IX-2, IX-3 and XI-4. The most intensive metasomatic dolomitization could be observed in the rock material of the boreholes XII-3, XII-5 and XIII-4 as well as in the lower part of boring XI-2.

The thermal spring activity and the metasomatic dolomitization, respectively, on the Nagyszál — taking into consideration the data found in the literature — extended from the Miocene to the Pleistocene.

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LETTERS OF THE WORKING GROUP ON MANGANESE FORMATIONS OF THE INTERNATIONAL ASSOCIATION ON THE GENESIS OF ORE DEPOSITS

PROVISIONAL PROGRAMME OF THE WORKING GROUP ON MANGANESE FORMATIONS SPONSORED BY THE INTERNATIONAL ASSOCIATION ON THE GENESIS OF ORE DEPOSITS (IAGOD)

The chief task of the Working Group on Manganese Formations (WGMF) established during the 2nd Symposium of the IAGOD held in 1967 in St. Andrews (Scotland) is to promote the exchange of scientific information, creative connections among scientists dealing with different aspects of geology, mineralogy and geochemistry of the manganese formations.

The Provisional Chairman of the WGMF is Professor GYULA GRASSELLY (Dept. of Mineralogy, Geochemistry and Petrography of the Attila József University at Szeged, Táncsics M. u. 2., Szeged, Hungary) appointed for this work by the Council of the IAGOD.

During his visit to Moscow (October 1969) the Provisional Chairman asked for the kind assistance of IGOR M. VARENTSOV (Geological Institute of the Academy of Sciences of the USSR, Pizhevski pereulok 7, Moscow Zh-17, USSR) to participate in the organizational work as Provisional Secretary of the WGMF.

Professor GRASSELLY and DR. VARENTSOV — national representative of the USSR in the WGMF — had thoroughly discussed the programme and future activity of the WGMF. The outlines and main points of the programme have been discussed with Professor VLADIMIR I. SMIRNOV, member of the Academy of Sciences of the USSR, Vice-President of the IAGOD.

Participation of the WGMF in the IMA-IAGOD Meeting

The statutory and the first scientific session of the Working Group on Manganese Formations of the IAGOD will be held during the IMA-IAGOD Meeting '70 (August 28—September 2, Tokyo—Kyoto, Japan). For the statutory meeting, discussion of organizational questions the Organizing Committee of the IMA-IAGOD Meeting '70 provided fixed time (August 28, Tokyo), and another time for the scientific session (September 2, Kyoto). See time-table of the Second Circular of the IMA-IAGOD Meeting.

The Working Group on Manganese Formations calls the attention of specialists in manganese ore researches to the session of the Working Group during the IMA-IAGOD Meeting in 1970, in Japan. National representatives are asked to promote that manganese ore research in their countries be worthily represented by papers submitted and lectures delivered at the session.

In case of any questions, please turn to the Provisional Chairman.

A provisional list of the staff of the Working Group on Manganese Formations of the IAGOD as well as a programme of the activity of the WGMF should be confirmed by the statutory meeting mentioned above.

The national representatives of the Working Group listed below were requested by the Provisional Chairman to undertake this task and they will be confirmed by the First Working Group Meeting, which will also elect the Chairman and Secretary.

Staff of the Working Group

GYULA GRASSELLY, professor, Head of the Dept. of Mineralogy, Geochemistry and Petrography, Attila József University, Szeged, Hungary, national representative of Hungary, Provisional Chairman.

IGOR M. VARENTSOV, senior scientific worker, Geological Institute of the Academy of Sciences of the USSR, Moscow, national representative of the USSR, Provisional Secretary.

SUPRIYA ROY, reader in Geology, Dept. of Geology, Jadavpur University, Calcutta, India, national representative of India.

JOHN VAN N. DORR II, research geologist, Geological Survey, Washington, U.S.A., national representative of the U.S.A.

LUBOR ŽAK, docent, Dept. of Mineralogy, Charles University, Prague, Czechoslovakia, national representative of Czechoslovakia.

LUBOMIR VASSILEFF, Geological Institute of the Academy of Sciences of Bulgaria, Sofia, national representative of Bulgaria.

VOJISLAV VUJANOVIČ, chief geologist, Inst. of Nuclear and Other Raw Materials, Belgrade, national representative of Yugoslavia.

Besides the above mentioned national representatives it would be desirable to greet among the members of the Working Group representatives of countries of Africa, South America, Australia, Japan and other countries interested in a comprehensive research of manganese formations. It is the duty of the Provisional Chairman to promote connections with the help of national representatives.

Publication of papers

Papers to be delivered in the IMA-IAGOD Meeting '70 will be published by the Organizing Committee (see Second Circular of the IMA-IAGOD Meeting '70 Tokyo—Kyoto). The date by which the abstracts (not more than 250 words) must be submitted to the Organizing Committee of the Meeting (Science Council of Japan, Ueno Park, Tokyo 110, Japan) is **before 31 December 1969** and papers **not later than 15 March 1970.**

The „semi-official" journal of the Working Group on Manganese Formations is Acta Mineralogica-Petrographica Universitatis Szegediensis, Szeged, Hungary, published yearly at the end of the year. It is planned to publish in this journal annual reviews of national representatives on manganese researches or summaries of papers published in different countries, concerning problems of manganese ore research. It is also possible to publish comprehensive papers on different aspects of geology and geochemistry of manganese formations.

Exchange of standard manganese mineral specimens

The WGMF thinks that it would be advisable if each researcher or at least the national representatives of the respective countries had a standard collection of well defined and tipical manganese mineral and ore specimens, respectively. Such a collection could facilitate the mutual understanding and interpretation as well as control of basic problems of mineralogy and geochemistry of manganese ores. Therefore, during the session of the WGMF in 1970 in Tokyo and Kyoto it would be desirable that the national representatives discussed the possibilities of preparation of sets of standard samples of manganese ores and minerals from the most profoundly investigated deposits of their countries. The exchange of such standard collections among national representatives will serve well as comparative material.

Preparation of the 2nd International Symposium on Manganese Ores

During the WGMF session in 1970 in Tokyo and Kyoto, besides the problems mentioned above it is also planned to discuss and begin the preparation of the programme of the 2nd International Symposium on Manganese Ores during the 24th International Geological Congress to be held in Montreal, Canada, 1972.

It seems advisable to discuss possible topics of the 2nd International Symposium on Manganese Ores outlined below. The topics will be coordinated by the experts denoted. Their duties include preparation of materials for future publication in the Transactions of the Symposium.

The main topics will be as follows:

I. General mineralogical and geochemical problems of the manganese ores (GY. GRASSELLY).

II. Manganese ores in volcano-sedimentary formations (JOHN VAN N. DORR II).

III. Manganese ores of ancient formations (SUPRIYA ROY).

IV. Processes of manganese ore formation. Manganese in natural waters (IGOR M. VARENTSOV).

The national representatives of the WGMF and other researchers interested in the activity of the Working Group and manganese ore researches may send their comments and proposals concerning the preparation and topics of the 2nd International Symposium on Manganese Ores 1972, Canada to the Provisional Chairman of the WGMF **not later than 30 June 1970.** Comments and proposals concerning the future activity of the Working Group on Manganese Formations are requested to be send at the same address by the date given.

Preparation of an international monography on manganese ores

During the WGMF session in 1970 in Tokyo and Kyoto it is planned to consider possibilities of preparing materials for a collective international monography on manganese ores. It may include the following chapters and topics, respectively.

I. Methods of investigation of manganese ores

1. Chemical analysis including general chemical analysis, rapid analysis, determination of minor elements by spectral analysis as well as methods of X-ray spectral analysis, neutron activation analysis
2. Ore microscopy of manganese ores
3. Electron microscopy
4. Infra-red spectroscopy
5. X-ray analysis, X-ray diffraction
6. Magnetometric measurements
7. Differential thermal analysis, thermogravimetry
8. Review of modern methods of physical analysis of solids applied to manganese ores

In the treatment of methods summary of available data, their critical review and presentation is included.

II. Geology, mineralogy and geochemistry of manganese ores
   1. Classification of manganese deposits
   2. Geology of the main types of manganese deposits
   3. Specific mineralogical and geochemical features of the different types of manganese deposits
   4. Processes of formation of manganese ores
III. Concise review of practical use of manganese ores
IV. Bibliography

In the course of the IMA-IAGOD Meeting to be held in 1970, in Japan, it will be possible to elaborate the above outlined draft in details, considering remarks sent in written form to the Working Group as well as verbal amendments done at the session. Furthermore, appropriate groups undertaking the task to elaborate certain topics could be formed during the session. The Working Group welcomes any help promoting this work by researchers interested in the programme of the Working Group and in manganese ore researches, respectively.

To solve the complicated scientific problems the WGMF recommends all the specialists to address their national representatives of WGMF for assistance in establishing scientific connections with researchers working in the particular narrow field of manganese ore researches.

All the above stated items are of preliminary character. They might be done more comprehensive and precise after discussion at the session of WGMF.

IGOR M. VARENTSOV
Provisional Secretary of the WGMF

PROF. GYULA GRASSELLY
Provisional Chairman of the WGMF

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